UNIVERSITY OF ALBERTA

LATE PLEISTOCENE STRATIGRAPHY AND SEDIMENTOLOGY OF COARSE CLASTIC DEPOSITS IN THE CENTRAL CANADIAN ROCKY MOUNTAINS, JASPER

BY

©

VICTOR M. LEVSON

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled Late Pleistocene Stratigraphy and Sedimentology of Coarse Clastic Deposits in the Central Canadian Rocky Mountains, Jasper submitted by Victor M. Levson in partial fulfillment of the requirements for the degree of Doctor of Philosophy.

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DEDICATION

This thesis is dedicated to all my family, particularly my parents, Arthur and Thelma Levson, my wife Barbara Levson, and my children Timothy, Sarah, Mary-Anne and Rebecca Levson, whose love, support and many sacrifices have made this work possible. I will do my best to make up for all those hours of missed family time. I love you all very much!

ABSTRACT

Six informal lithostratigraphic units are recognized from newly mapped exposures of Late Quaternary deposits in the Jasper region. The oldest deposits consist mainly of locally derived, coarse-grained gravelly diamicton interpreted as large volume, episodic, unchannelled, debris flow deposits. The inferred paleo-environment is a cool temperate, alpine to subalpine alluvial fan setting. Radiocarbon dates indicate that fan sedimentation began prior to 48 ka and continued throughout the Middle Wisconsinan. A unit of stratified gravels and sands interpreted as braided stream glaciofluvial deposits gradationally overlie the paleofan sequences and are distinguished from them by better stratification, sorting, clast roundness and relatively high percentages of distally-derived clasts. Wood from this unit near Jasper town site indicates deposition about 29 ka. Glaciolacustrine sediments overlying this unit in several Front Range valleys, reflect the onset of glaciation in the Athabasca River valley and associated ice-damming of tributary valley streams. Both Rocky Mountain and Cordilleran glaciers advanced through the area, depositing a complex sequence of ice-marginal sediments, basal tills and supraglacial deposits. Several lines of evidence suggest that at the Late Wisconsinan maximum, glaciers flowed out of the Athabasca valley into the conjectural 'ice free' corridor and were deflected southeasterly along the mountain front by the Laurentide ice sheet.

During deglaciation, sedimentation in the Athabasca valley was dominated by ice-marginal deposition of glaciofluvial sands and gravels and by paraglacial debris flows. Large medial moraines and debris septa that formed at the confluence of main valley glaciers, such as occupied the Miette and Athabasca valleys, probably led to stagnant ice in the valley centre and associated kame terrace development along the valley sides. A radiocarbon date on shells from a small ice-marginal lake near Pocahontas indicates that glaciers there were in retreat by about 12 ka and alpine glaciers in the region were at or near their present limits by 10 ka.

Sedimentologic studies of alluvial fan deposits in the area indicate that fast moving, noncohesive debris flows are prevalent and that the bedrock geology of the fan source areas exerts a strong control on sedimentation patterns. The conclusion that high energy, low frequency, debris flow events are prevalent in the Main Ranges, whereas lower energy, higher frequency events are typical of the Front Ranges is of relevance to modern hazard studies and mitigation efforts in the Canadian Rocky Mountains.

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CHAPTER 1 - INTRODUCTION

The geology of Late Pleistocene deposits in many parts of the Canadian Cordillera is poorly understood due largely to the scarcity of stratigraphic and detailed sedimentologic data. The genesis of clastic deposits in the subsurface are particularly enigmatic, not only because good exposures are relative rare but also because these deposits may have no geomorphic expression from which genetic inferences can be made. This is particularly true when they stratigraphically underlie sediments of one or more Pleistocene glaciation ('sub-till deposits') and interpretations on deposit genesis therefore must rely mainly on sedimentologic data. Determining the origin of coarse clastic deposits is further complicated because these sediments may originate in a wide variety of depositional environments including: fluvial settings; glaciofluvial environments; alluvial fan, delta and fan-delta environments; colluvial settings (including creep, talus, debris flows and slides); and glacial environments (supraglacial, englacial and subglacial settings). Since deposits from many of these environments have similar sedimentary characteristics, genetic interpretations are commonly equivocal. In addition, due to difficulties in studying and modeling the behavior of high energy flows capable of transporting large clasts, both in natural flows and in laboratory flume studies, genetic interpretations of coarse clastic deposits must be inferred from their sedimentary characteristics. This is especially the case for subsurface sediments where geomorphological data or other indirect evidence of origin such as paleontological information are lacking.

These comments suggest that there is a need for detailed field studies of coarse clastic deposits originating in a variety of different depositional environments. Studies of this type that compare deposits from a wide range of environments and geologic settings are lacking. This thesis is an investigation of Late Pleistocene coarse clastic deposits from varied depositional settings in the Jasper National Park area where few studies of this type have been conducted. Detailed stratigraphic and sedimentologic investigations are provided for a number of previously undescribed sites in the study region. Alluvial fans and gravelly braided stream settings dominated sedimentation in most of these areas during interglacial and preglacial times. The sedimentary record of the transition from interglacial to glacial environments at the onset of the Late Wisconsinan is also observed at some sites and glacigenic deposits from this glacial event are present in most areas. Since coarse clastic deposits can occur in many different depositional environments, emphasis is placed in this thesis on identifying the sedimentologic criteria needed to differentiate their genesis. The stratigraphic relationships of the sub-till deposits to the overlying glacial sequences are described, as are the sedimentologic characteristics of the deposits themselves. This data is used to identify criteria for distinguishing glacial from non-glacial deposits.

In addition to investigating the genetic origin of Late Pleistocene deposits in the Rocky Mountains this research provides some new chronostratigraphic data for these deposits. Chronostratigraphic data are presently limited for several reasons. Firstly, coarse-textured clastics are especially difficult to date because of the low preservation potential of tephra and organic materials in high energy depositional environments. Secondly, sub-till deposits typically are beyond the range of radiocarbon dating (~40 000 years). Dating problems with coarse clastics are exacerbated in the eastern parts of the Cordillera where materials suitable for radiometric dating are relatively rare. This is in contrast to more tectonically active regions of the Cordillera where extensive volcanism has resulted in interbedded tephra, basalt and other volcanic materials suitable for K-Ar dating. In addition, the relatively low biological productivity of the eastern Cordillera is not conducive to the occurrence, in clastic deposits, of wood, shells and other organic materials from which ^{14}C dates can be obtained (Clague, 1981).

The sub-till Cenozoic deposits described in this thesis occur in a number of different geological and glaciological settings. The study sites discussed represent examples of depositional styles in each of the two main morphogeotectonic belts in the eastern Rocky Mountains: the Main Ranges and Front Ranges. The Jasper area is also located within the influence of both the Cordilleran ice sheet and local montane valley glaciers and therefore is ideal for the study of relationships between the two glacier complexes. The influence of regional geological and glaciological differences on the deposition and preservation of Pleistocene deposits in the area is emphasized throughout this thesis.

The four chapters that follow this introduction focus on different but related aspects of the stratigraphy and sedimentology of Late Pleistocene glacial and nonglacial deposits in the Jasper region. A stratigraphic overview of Quaternary deposits throughout the study region is first provided in Chapter 2. Emphasis is placed in this chapter on sub-till deposits, as glacigenic deposits in the region were the focus of previous studies (Levson, 1986; Levson and Rutter, 1986, 1989a). Chapter 3 provides a detailed description of the sedimentology of a succession of buried alluvial fan deposits that comprise the oldest lithostratigraphic unit in the area. Differences in sedimentation in the paleofans between the Rocky Mountain Main Ranges and the Front Ranges are discussed. The fourth chapter is a discussion of a complex sequence of glacial and glaciofluvial sediments in the Jasper town site area that reflect a complete cycle of glaciation in the Late Wisconsinan. Geomorphic, stratigraphic, and sedimentologic data obtained from Quaternary sediments in the area are used to reconstruct depositional environments in the vicinity of two former confluent valley glaciers. Emphasis is placed on the origin of proximal and distal glaciofluvial gravels deposited during both advance and retreat phases.

Regional implications of the Jasper Pleistocene stratigraphy are discussed in Chapter 5. In particular, the implications of new constraints on the timing of glaciation in Jasper National Park are investigated. The constraints are provided mainly by a radiocarbon date of 29,100 ± 560 years B.P. (GSC 3792) obtained on wood in gravels underlying till (Levson and Rutter, 1989b) and a shell date of 11,900 ± 120 years B.P. (GSC 3885) from ice-marginal lake sediments (Levson *et al.*, 1989). These dates and the associated stratigraphic relationships indicate an extensive Late Wisconsinan glaciation in the Jasper National Park region (*c.f.* Chapters 2 and 4). One ramification of this conclusion is that glaciers moving down the Athabasca River valley in the Late Wisconsinan flowed out of the study area into a region, known as the 'ice-free' corridor, where the existence of Late Wisconsinan glaciers has long been disputed. Chapter 5 also includes a discussion of other evidence suggesting that the ice-free corridor was glaciated during the Late Wisconsinan.

This research is intended to expand existing knowledge of the stratigraphy and sedimentology of Quaternary deposits in the Canadian Cordillera and improve current understanding of the genesis of coarse clastic deposits. The identification of criteria that help distinguish glacial from non-glacial clastic deposits is useful for modeling past climatic and environmental changes. Stratigraphic and sedimentologic studies of Pleistocene gravelly sediments are also economically important for several reasons. In many areas of the Canadian Cordillera interglacial and preglacial Cenozoic gravel sequences host important gold-bearing placer deposits (Levson and Giles, 1993). A good comprehension of the genesis of these deposits is required to help locate new buried deposits. Glaciofluvial and alluvial deposits also provide important sources of aggregate and since these coarse deposits are typically highly permeable they also may be important aquifers (Edwards, 1989; Halstead, 1991). An understanding of their stratigraphy and sedimentology is often required for sand and gravel assessments, and groundwater resource and contamination studies. Finally, detailed investigations of high energy clastic deposits may have important implications for slope hazard research in steep mountainous areas (Clague, 1991).

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CHAPTER 2 - PLEISTOCENE STRATIGRAPHY OF THE ATHABASCA RIVER VALLEY REGION, ROCKY MOUNTAINS, JASPER NATIONAL PARK, CANADA[1]

INTRODUCTION

The preservation of Quaternary sediments in alpine regions of the Canadian Cordillera is generally poor because of repeated erosion by Pleistocene glaciers and Holocene mass wasting. Large valleys, however, acted as sediment accumulation areas during both glacial and nonglacial times. The resultant valley-fill sequences, although typically incomplete due to erosional unconformities, provide the best record of Pleistocene events in the Cordillera (Clague, 1991). The purpose of this chapter is to present data from a part of the Rocky Mountains where the Pleistocene stratigraphy is relatively poorly known. The area centres on the part of Jasper National Park that is dissected by the Athabasca River valley (Fig. 2-1a).

In the Canadian Rocky Mountains, investigations of relatively thick valley-fill sequences have been restricted almost exclusively to the Rocky Mountain Trench (*e.g.* Clague, 1975; Rutter, 1976, 1977; Fulton and Archard, 1985) but there are few descriptions from sites within the Rocky Mountains east of the trench. Previous studies of the Quaternary stratigraphy of the Rockies were undertaken in the Peace River area approximately 450 km northwest of Jasper town site (Mathews, 1978, 1980) and along the North Saskatchewan River valley (McPherson, 1970) and the Bow River valley (Rutter, 1972), 150 and 250 km southwest of Jasper, respectively.

Pleistocene sediments are exposed at a number of localities in the study area (Figs. 2-1a and 2-1b) and reveal a complex sequence of alluvial fan, glaciofluvial, glaciolacustrine and glacial deposits. The stratigraphic record is best preserved within the Athabasca River valley and major tributaries such as the Miette, Snake Indian and Rocky River valleys. Large stratigraphic sections also occur at the mouths of smaller tributary valleys such as the Portal Creek valley (Levson and Rutter, 1986). The preservation of Pleistocene sediments in the upper reaches of these valleys is restricted mainly to tills deposited during the last glaciation.

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[1] *A version of this chapter has been accepted for publication: Levson and Rutter, 1995. Géographie physique et Quaternaire, Volume 49, Number 3.*

Study Area

The Rocky Mountains consist of southwest-dipping thrust sheets, of mainly Paleozoic and Precambrian carbonate and clastic strata, formed during Late Cretaceous-Early Tertiary orogenic events. The southeast striking Pyramid Thrust, located directly east of Jasper town site, separates the Rockies into the Main Ranges, to the southwest, and the Front Ranges, to the northeast. The Front Ranges are composed of steep, southwesterly dipping, thrusted carbonate, shale and sandstone strata from upper Cambrian to Jurassic in age. The Main Ranges consist of thick, relatively flat lying or gently dipping, thrust blocks with broad folds and late, extensional, normal faults. The rocks are quartzose sandstones, shales, siltstones and carbonates of Cambrian and Precambrian age (Price and Mountjoy, 1970; Price *et al.,* 1977; Mountjoy and Price, 1985). Relief in the Front and Main Ranges is high, commonly in excess of 1500 m. Mount Robson, at an elevation of 3955 m in the western Main Ranges just west of Jasper National Park (Fig. 2-1a), is the highest peak in the Canadian Rockies.

The surficial geology of the study area is shown in Figure 2-1c. The Athabasca River valley and other large river valleys in the area are characterized by gravelly and sandy, floodplain alluvium as well as fluvial and glaciofluvial terrace deposits (units Ap and F^G on Fig. 2-1c). Gravelly alluvial fans (Af) encroach on the valley floors in many areas. Upland areas are characterized by thin (less than 2 m) colluvium (Cv) and exposed rock whereas lower valley sides are blanketed or veneered by morainal debris (Mb and Mv). Areas of thick colluvium (Cb) include mainly rock avalanche and large talus deposits. Units of organic deposits (O) and eolian sand (E), large enough to show on Figure 2-1c, are restricted mainly to the area west of Jasper and to the Jasper Lake region, respectively.

The size and orientation of major valleys in the area are commonly controlled by regional bedrock structures. The Athabasca River flows northwesterly along the strike of the Main Ranges, from the Columbia Icefields (Figs. 2-1a and 2-1c) to the Jasper town site area where it then flows northeastwards across the strike of the Front Ranges. In the Front Ranges, tributary valleys of the Athabasca River, such as the Snake Indian and Rocky River valleys, occupy areas underlain by relatively weak shales and sandstones between mountain ranges dominated by resistant Paleozoic carbonates. These valleys tend to follow the southeasterly structural trend and they are relatively large compared to most tributary valleys of the Athabasca River in the Main Ranges. The Miette River valley, crossing the strike of the Main Ranges west of Jasper town site, is one exception (Figs. 2-1a and 2-1c).

Seasonal climatic variability in the study area is high. Mean January and July temperatures at Jasper town site are approximately -13°C and +15°C, respectively (Atmospheric Environment Service, 1982). Temperature ranges throughout the year are broad with snow falling at any time of year, especially at high altitudes. Approximately 400 millimetres of precipitation falls annually, including about 150 centimetres of snow. Climatic variations are also large over short distances due to the effects of high relief. Precipitation along the Continental Divide, for example, is much higher than in the valley bottoms and temperatures are lower, resulting in the development of extensive icefields, alpine glaciers and rock glaciers (Fig. 2-1c). The climate of the study area and related discussions of the region's vegetative and soils characteristics were provided by Holland and Coen (1983) and Levson *et al.* (1989).

Previous Quaternary Studies

The surficial geology of the study area was mapped by Reimchen (1976) and as part of a biophysical mapping program by Holland and Coen (1982). A lithofacies classification of glacial deposits in Jasper National Park was provided by Levson (1986) and revised by Levson and Rutter (1989a). The Quaternary stratigraphy of the Portal Creek area (Fig. 2-1a) was described by Levson and Rutter (1986) as was the Quaternary history of the Jasper town site area (Levson and Rutter, 1989b). Roed (1975) investigated the Quaternary geology of the Athabasca River valley region immediately east of Jasper Park in the Hinton-Edson area and Roed *et al.*, (1967) discussed the origin of the Athabasca Valley erratics train. Holocene glaciations in the Jasper region have been investigated by numerous authors including Luckman and Osborn (1979), Kearney and Luckman (1981, 1983a, b, 1987), Beaudoin (1984, 1986), Luckman (1986, 1988), Luckman and Kearney (1986) and Osborn and Luckman (1988). Levson *et al.* (1989) provided a review of the Quaternary geology and paleoecology of the region as well as a number of detailed descriptions of sites where Late Wisconsinan and Holocene paleoenvironmental and geological data have been collected.

LITHOSTRATIGRAPHY OF THE ATHABASCA VALLEY

The locations of measured sections in the study area are shown on Figures 2-1a and 2-1b and composite stratigraphic columns from several main areas are given in Figure 2-2. Five main lithostratigraphic units are recognized in the region. From oldest to youngest these units consist of massive, clast-supported diamicton and poorly sorted gravels (Unit 1), stratified gravels and sands (Unit 2), parallel laminated silts, clays and fine sands (Unit 3), matrix-supported diamicton (Unit 4) and a relatively thin upper sequence of clay, silt, sand, gravel and diamicton (Units 5 and 6). Pronounced unconformities occur at the base of Units 4 and 5. Descriptions and interpretations of each unit follow. Correlations of units between sites are based mainly on stratigraphic position, deposit type, and lithologic composition.

Unit 1 - Massive, Clast-Supported Diamicton and Gravel (Paleoalluvial Fan Deposits)

The lowermost succession of deposits at the described localities consists of up to 50 metres of mainly clast-supported, gravelly diamicton (Unit 1, Fig. 2-2 and Fig. 2-3a). Interbedded massive and cross-stratified gravels and minor sands also occur in this unit. (Gravel and diamicton are distinguished by the presence of mud in the latter; see also Chapter 3). Diamicton beds containing cobble to boulder-sized clasts dominate and are usually more poorly sorted than pebbly diamicton beds. They are generally poorly sorted, massive or crudely bedded and up to several metres thick. Clasts are subrounded to angular (Fig. 2-3b). Large pebble to cobble gravel interbeds are locally common, especially in the Front Ranges. Well to moderately well sorted gravel beds exhibiting horizontal, planar and trough cross-stratification are less common as are sandy strata and discontinuous interbeds and lenses of open work gravel. Beds dip generally about 5° to 8° but up to 11°, towards the valley centre (Fig. 2-3a). Clasts of local, tributary basin origin dominate (Fig. 2-4). This unit contains mostly cobble to boulder-sized clasts in the Main Ranges and pebbles to cobbles in Front Range sections. Deposits of this unit are described and discussed in detail in chapter 3.

Interpretation

An alluvial fan origin for these deposits is indicated by the consistent dip of strata towards the valley centre and by the dominance of clasts derived from tributary drainage basins on the adjacent valley sides. Also consistent with an alluvial fan origin is the crude stratification, coarse grain-size, high clast angularity and overall poor sorting. The dominance of massive to crudely bedded diamicton beds suggests that deposition occurred mainly from episodic debris and flood flows associated with fan development and to a lesser extent from intermittent fluvial activity (see also Chapter 3).

Erratic lithologies are rare or absent in the bulk of the fan sediments but erratics and glacially abraded clasts are abundant in diamicton units (see below) that stratigraphically overlie the fan sequences (Fig. 2-3a). This is believed to reflect the transition from non-glacial to glacial depositional environments. This transition is also locally shown by paleoflow indicators and provenance studies of clasts that indicate that a significant component of clasts in the uppermost part of some of the fan sequences are derived from upstream regions in the main valleys (Fig. 2-4, Unit 1-2 transition) rather than from tributary valleys. The local influx of clasts of main valley provenance probably reflects the influence of glaciofluvial activity during the advance of ice into the area.

Unit 2 - Stratified Gravels and Sands (Fluvial / Glaciofluvial Deposits)

The paleoalluvial fan deposits of Unit 1 are stratigraphically overlain by a complex, bedded gravel and sand sequence (Unit 2, Fig. 2-2) that is divided into two subunits. The lower subunit, interpreted as fluvial or distal glaciofluvial sediments, grades upwards into the upper subunit, interpreted as proximal braided stream deposits. The thickest exposures of the unit occur in the Pocahontas (Figs. 2-1b, 2-2f and 2-5a), Jasper town site (Fig. 2-6a) and Punchbowl Falls (Figs. 2-1b, 2-2g, 2-5b) areas. These deposits are generally found about 100 to 200 m higher than the modern rivers.

The gravels are mainly crudely to well stratified, well imbricated and moderately to well sorted. They are typically horizontally bedded (Fig. 2-6a) but planar and trough cross-bedding are also locally common (Fig. 2-5b). Clasts are mainly subrounded to well-rounded pebbles to small cobbles. In comparison to the underlying fan deposits these gravels have more clasts derived from main valley basins (Fig. 2-4). In the Front Ranges, for example, where quartzitic rocks indicate transport of clasts down major valleys from the Main Ranges, gravels in Unit 2 typically contain 20-30% quartzites compared to less than 5% in Unit 1. At a few sites such as Roche Miette the lithologic contrast is abrupt (Fig. 2-3b) whereas at most sites, such as Jacques Creek and Pocahontas, transitional gravels, occurring between Units 1 and 2, contain intermediate quantities of quartzite (5-20%). Paleoflow directions also indicate deposition in main valley fluvial systems that generally flowed parallel to the present day Athabasca River. A conformable contact between gravels in Unit 2 and the underlying paleofan sequences is locally indicated by interbedding of diamicton, of probable alluvial fan origin, in the lower part of the unit and by the typically gradual, increase in abundance of distally derived rocks between Units 1 and 2.

Lower Subunit

Sediments in the lower part of this unit are dominated by horizontally bedded gravels, sands and minor silts (Fig. 2-6a). Trough and planar cross-bedding are restricted mainly to sandy strata or thin pebble gravel beds. Gravels are imbricated, poorly to well sorted, clast-supported and generally have a medium to coarse sand matrix; beds average 40 to 80 cm in thickness. Clasts are mainly pebble to cobble-sized and subrounded to well rounded. Planar cross-stratified sand beds up to 4 cm thick occur sporadically within the gravels. The grain-size distribution of most individual beds are laterally uniform over tens of meters.

Fining-up sequences 3 to 15 m thick dominate this subunit (Fig. 2-6a). Clast-supported gravels at the base of the sequences exhibit large-scale, crude horizontal bedding. They grade upwards into matrix-supported, pebbly sand beds. Sandy strata, dominating the upper part of each cycle, show horizontal laminae, planar and trough cross-bedding and climbing-ripple cross-laminae. Massive sand beds, up to one meter thick, are also common. Dish structures occur locally in some, otherwise massive, sandy beds. Most fining-upwards cycles are capped by beds of sandy silt, a few centimeters thick, that may extend laterally for tens of meters before pinching out. They exhibit wavy laminae, load and injection structures, and may contain up to 50% granules and small pebbles. Some overall fining-up units are underlain by a coarsening-up sequence, up to 0.5 m thick.

Organic materials, including wood and gastropod and pelecypod shell fragments, occur throughout this subunit but they decrease in abundance towards the top. A radiocarbon date of 29,100 ± 560 years B.P. (GSC-3792) was obtained from wood in these deposits near Jasper town site at a depth of 17.9 m below the surface (1115 m asl).

Upper Subunit

Sediments in the upper subunit exhibit a much higher degree of sedimentologic variability than in the lower subunit. In addition to horizontal bedding, planar cross-bedded gravels and large trough cross-bedded sand and gravel lenses, up to several metres thick and tens of metres wide, are locally common (Figs. 2-5b and 2-6b). Coarse gravel commonly occurs at the base of these lenses and the overlying sediments generally fine upwards. Some coarsening-upward sequences occur in the upper part of this subunit. Scour-and-fill structures also increase in abundance towards the top of the subunit as does the coarseness of the lag gravels at the bottom of the scours. Diamicton intrabeds containing striated clasts and poorly sorted coarse gravels also occur in some areas in the upper part of this subunit (Figs. 2-2a, 2-2b and 2-2d). Strata locally dip steeply in various directions and show lateral and vertical variability in grain size and sorting both within and between beds. Faults and load and injection structures are common. In addition, bedding in the upper few metres of the gravels at some sites shows overturned folds and thrust faults (Fig. 2-2e). Deformation is most intense in the vicinity of large 'embedded' boulders (Fig. 2-5c) that commonly have faceted and striated upper surfaces.

Gravels of this subunit at Amber Mountain (Fig. 2-2a) are restricted to the valley side above a sequence of fan deposits of Unit 1 (Fig. 2-6b). The sands and gravels at this site are laterally interbedded with more poorly sorted gravels and diamicton beds.

Interpretation

Well developed pebble roundness, imbrication, cross-bedding and sorting in gravels of this unit indicate a fluvial depositional environment. The dominance of horizontally bedded gravel suggests deposition mainly in longitudinal bars. Crude horizontal stratification is typical of longitudinal bars in shallow, gravel-bed, braided streams (Smith, 1974; Hein and Walker, 1977). Similar deposits characterized by horizontally stratified gravels with thin interbedded sand lenses and showing ripple-drift cross-laminae and mud veneers, have been described in modern braided stream systems (Boothroyd and Ashley, 1975; Church and Gilbert, 1975).

Fining-upward sequences, dominant in the lower subunit, probably formed as a result of waning flow conditions subsequent to major flood events. Braided stream deposits characterized by fining-up gravel and sand cycles, formed under progressively decreasing energy levels in successive flood events have been described by several authors (Rust, 1972, 1978; Boothroyd and Ashley, 1975; Miall, 1977; Rust and Koster, 1984). Since no disconformities are present within each cycle, deposition may have occurred during single large flood events (*c.f.* Maizels, 1989). Cyclicity in the sedimentation pattern probably resulted from re-occupation of the channel during subsequent high flow events. Preservation of fine-grained sediments at the end of each cycle suggests that subsequent flows were initially depositional and increased in strength gradually. This is supported by a coarsening upwards of grains at the base of some otherwise fining-upwards sequences. Silt and clay beds capping the cycles probably were deposited by suspension from muddy waters in recently active channels cut-off by falling water levels.

Sediments in the lower subunit are believed to have been deposited in a distal braided stream environment. The lateral uniformity and continuity of these sediments supports this interpretation. The relatively high abundance of wood fragments, gastropods and pelecypod shells in the lower subunit suggests a more viable environment for flora and fauna, possibly also reflecting a greater distance from glaciers. In addition, parallel laminated mud facies, such as occur throughout the lower subunit, are more common in distal or inactive reaches of modern glaciofluvial streams, where preservation potential of muds is higher, than in proximal reaches (Rust, 1972). The lack of organic matter in the top of the lower subunit and throughout the upper subunit is consistent with this interpretation, and suggests that a progressive transition, from fluvial, or distal glaciofluvial, to more proximal glaciofluvial environments, occurred between the deposition of the lower and upper subunits.

The sedimentologic complexity of the upper subunit reflects a more varied depositional environment. In addition to longitudinal bar deposits, planar cross-bedded gravels interpreted as transverse bar deposits, occur in the upper subunit. Large trough-shaped units of fining-up channel-fill gravels and sands at some sites, indicate that well developed channels also occurred locally. The restriction of these gravels at Amber Mountain to sites high along the valley side (Fig 2-6b), suggests that they may have been deposited in channels along the margin of a glacier occupying the centre of the Athabasca valley. This confined flow probably resulted in the more pronounced channelization of the gravels in this area compared to most other sites. Coarse gravel concentrations at the base of some of these channels are interpreted as lag gravels resulting from winnowing and channel erosion.

Deposits in the upper subunit are inferred to have a more proximal glaciofluvial origin than gravels in the lower subunit. In the Jasper town site and Roche Miette areas (Figs. 2-1a and 2-1b), an upward increase in the number of cut and fill structures, erratic clasts, overall grain size of the unit and coarseness of lag gravels, as well as a vertical decrease in abundance of wood and shell fragments, are believed to reflect the increasing proximity of glaciers. Large variations in sediment sorting, grain-size distribution, clast size and shape, and paleocurrent direction in this subunit, also indicate proximity to a sediment source and highly variable flow conditions, typical of proximal ice-marginal environments (Banerjee and McDonald, 1975; Church and Gilbert, 1975; Boulton and Eyles, 1979). This interpretation is further supported by the local presence in the upper part of this subunit of diamicton intrabeds containing striated clasts of non-local lithology. These diamicton beds have characteristics similar to glacially-derived debris flows described in modern glacial environments (*e.g.* Lawson, 1979a) and they were probably deposited in ice-marginal streams of the advancing glaciers. Massive diamicton beds overlying the gravels at most sites (Fig. 2-2) are interpreted as tills deposited by these glaciers (see Unit 4 below). The presence of faceted, embedded and striated clasts as well as inferred glaciotectonic deformation structures such as overturned folds and thrust faults in the upper part of these gravels at some sections also indicates that they were eventually overridden by glacial ice (*e.g.* Figs. 2-2e, 2-5c).

Unit 3 - Parallel Laminated Silts, Clays and Fine Sands (Advance-Phase Glaciolacustrine Deposits)

In the lower reaches of several valleys tributary to the Athabasca valley such as the Jacques, Snake Indian and Rocky River valleys (Fig. 2-1a), glaciofluvial gravels in Unit 2 are overlain by horizontally laminated silts, clays and fine sands (Figs. 2-2 and 2-7). These sediments are very well sorted, rhythmically bedded and contain rare isolated clasts, some of which are striated. The unit commonly coarsens upwards from dominantly, parallel laminated, silts and clays at the base (Fig. 2-7a) to mainly trough cross-bedded and ripple-bedded silts and sands near the top (Fig. 2-7b). Diamicton intrabeds, locally exhibiting flow folds as well as load and injection structures, and some sand and gravel lenses also occur in the upper part of the unit (Fig. 2-7c). The lower contact of the unit is generally sharp and planar (Fig. 2-7a).

The sediments are interpreted as glaciolacustrine deposits. Parallel laminated, silts and clays at the base of the unit represent relatively quiet water deposits whereas cross-bedded sands are inferred to be more proximal glaciolacustrine sediments, deposited by streams discharging into the lake. Frequent interbedding of the sands with quiet water silts and clays suggests that the former are more likely to be glaciolacustrine than glaciofluvial in origin. Diamicton intrabeds common in the upper part of the unit are interpreted as subaqueous debris flow deposits, also because of their association with laminated silts and clays. Vertical facies changes in the unit are equivalent to the distal to proximal, glaciolacustrine facies sequence described by Ashley (1989) and probably reflect the increased proximity and eventual advance of glaciers into the lakes as glaciation in the region progressed. Similar facies associations also have been described from other glacial lakes in mountain regions (*e.g.* Shaw and Archer, 1979; Gilbert and Shaw, 1981; Ashley *et al.*, 1982; Ashley, 1989).

Ice-dammed lakes were a characteristic feature of tributary valleys in the study area at the onset of the last glacial period, particularly in Front Range valleys that remained ice free during the early phases of glaciation (Levson, 1990). The regional distribution and stratigraphic position of this unit suggests that deposition began when the drainage in tributary valleys was suddenly dammed, presumably by glaciers in adjacent trunk valleys. This interpretation is supported by the sharp lower contact of parallel laminated silts and clays conformably overlying coarse gravels, indicating a rapid transition from glaciofluvial to glaciolacustrine sedimentation.

Unit 4 - Matrix-Supported Diamicton Sequences (Glacial Sediments)

Diamicton facies in the area have been described in detail by Levson and Rutter (1989a). Three of the main facies associations recognized are: 1) fine-grained, massive, compact diamicton with rare sand lenses (upper unit in Figs. 2-6a and 2-6b), 2) silty to sandy, very poorly or poorly sorted diamicton, usually with interstratified gravels and sands or fine sediments, and 3) loose, sandy or gravelly diamicton containing numerous angular clasts. The first facies association dominates Unit 4, whereas the latter two are restricted mainly to the upper part of the unit.

Units 1 to 3 at all of the study sites are stratigraphically overlain by up to 20 metres of dominantly massive, matrix-supported diamicton (Unit 4, Fig. 2-2; Figs. 2-3a, 2-5a, 2-6a, b). The diamicton is typically unsorted, dense and has a fine-grained matrix. Striated and faceted clasts are common and most are subangular to rounded. Up to 85% of the clasts are the same lithology as the underlying or local bedrock. Pebble fabric analyses of the long axes of clasts indicate a strong unimodal preferred orientation parallel to the valley (*e.g.* S_1=0.74 at the Pocahontas section, Figs. 2-2f and 2-9a). The basal contact of the diamicton is generally sharp and erosional, truncating bedding in underlying sand and gravel units (Figs. 2-2f, 2-3a, 2-5a, 2-6a and 2-6b). Clasts at the base of this unit locally are embedded in underlying deposits and commonly have faceted and striated upper surfaces with both striae and clast long-axes showing preferred orientations parallel to the valley (Fig. 2-9a). Shear structures and compressively deformed bedding in underlying sediments are locally apparent. Lenses of sorted sand and gravel, typically plano-convex or circular in cross-section, are rare. Stratification within the diamicton is also rare and occurs as diffuse beds with slight textural differences that commonly drape underlying boulders (see Fig. 8 in Levson and Rutter, 1989a). The diamicton surface at some sites is fluted.

Diamicton beds with weak stratification, relatively high clast contents, silty to sandy matrix textures and poorly developed fabrics commonly cap, and in a few cases underlie, denser, massive diamicton sequences (Fig. 2-2). Pebble fabric data from these diamicton beds display weak trends (*e.g.* S_1=0.55 in the upper most diamicton unit at the Pocahontas section, Figs. 2-2f and 2-9b), commonly oblique or transverse to ice-flow direction. Diamicton beds dominate but locally present are, less than 1 metre thick, sand and gravel lenses that are typically trough-shaped, moderately to well sorted and stratified. Interstratified lenses and beds of gravels, sands and silts commonly are faulted, folded or show other evidence of post-depositional deformation. Shear structures and thrust faults locally occur in these strata near the base of the unit.

Sandy, poorly consolidated, diamicton, containing angular clasts up to boulder size, locally occurs on the surface in the study area and typically overlies massive, compact diamicton or bedrock. Diamicton beds are often interbedded with sands and gravels and are usually poorly exposed due to their poor compaction, sandy texture and thinness (generally less than one meter thick). The diamictons occur sporadically and locally are recognized only by the presence of scattered, angular, low to medium grade, metamorphic erratics. These erratics are commonly mica-garnet schists and biotite-muscovite schists that occur mainly on the surface of diamicton units and on surrounding bedrock exposures, most commonly in the Front Ranges up to an elevation of about 1400 metres. Source areas for these metamorphic erratics occur only west of the Continental Divide (Roed *et al.*, 1967).

Interpretation

Dense, fine-grained, massive diamicton beds in Unit 4 are interpreted as basal tills deposited by both lodgment and meltout processes. The main characteristics of the diamictons, including fine matrix textures, over consolidation and abundant glacially abraded clasts, are typical of basal tills (Boulton, 1976; Kruger, 1979; Haldorsen, 1982; Dreimanis, 1989). This interpretation is supported by the preferred orientations of clast long-axes which parallel former ice-flow directions (as determined primarily by valley trend). Sedimentary structures indicating lodgment processes include numerous faceted and embedded clasts, shear structures and strong valley-parallel pebble fabric trends (Boulton, 1976, 1978; Kruger, 1984). The local presence of 'relict' englacial features, such as rare, plano-convex or circular sand and gravel lenses, and diamicton strata that drape large clasts, suggests that deposition by meltout processes also occurred (Haldorsen and Shaw, 1982; Shaw, 1982; Levson and Rutter, 1989a). Erosional basal contacts probably formed at the sliding base of overriding ice, and fluted topography also reflects glacial cover over these deposits. Evidence for subglacial fluvial erosion is also present and includes deep and narrow incision of Unit 2 sediments in the valley center (see Chapter 4) and a relatively low concentration of rounded quartzite clasts from Unit 2 in the tills, as would be expected if only ice was responsible for erosion of the underlying gravels. In addition, the locally non-erosive nature of the glaciers is evidenced by preservation of thick units of unconsolidated sediments as well as proximal glaciofluvial gravels and ice-marginal deposits (Units 1 to 3) under till at a number of locations (*e.g.* Figure 2-2a, c, d and e). Tills at all of the sites were deposited mainly by Athabasca Valley ice but the influence of tributary-valley glaciers at valley junctions is indicated by stratigraphically complex diamicton sequences of variable provenance (*e.g.* Levson and Rutter, 1986).

Poorly sorted, diamicton deposits in this unit with weak stratification, chaotic fabrics, relatively high clast contents, and interbedded sand and gravel lenses are typical of debris flow deposits in glacial environments (Marcussen, 1975; Lawson, 1979a, b, 1981a, b, 1982, 1989; Jong and Rappol, 1983; Dreimanis, 1989). Stratification, relatively high clast contents and sandy matrix textures in these diamictons are interpreted to reflect the influence of gravity and water reworking of the debris flow deposits. Diamicton beds rich in gravel and sand may have undergone several cycles of resedimentation resulting in the removal of silts and clays by winnowing (*c.f.* Lawson, 1979a, 1981a, b). Lenses and beds of sorted material are interpreted as ice-proximal glaciofluvial sediments deposited between debris flow events. Lawson (1979a, 1989) found that diamictons deposited by individual flows in modern ice-marginal environments were often separated by layers of glaciofluvial sand and silt. Shear structures and thrust faults in these strata near the base of the unit are interpreted as glaciotectonic features.

Diamicton beds in the area containing distally derived (>50 km from source) metamorphic erratics, probably originated as supraglacially transported debris. The metamorphic erratics are lithologically similar to, and interpreted to be part of, the supraglacially transported Athabasca Valley erratics train, first described by Roed *et al.* (1967) from the area east of Jasper National Park. The sandy matrices, poor consolidation, high clast angularity and large modal clast-size of the diamictons and their common superposition over deposits interpreted as basal or englacial meltout tills (Levson and Rutter, 1989a), suggests that they were supraglacially derived. Similar characteristics have been described for supraglacial tills from both ancient and modern glacial deposits (Sharp, 1949; Drake, 1971; Stewart and MacClintock, 1971; Boulton and Eyles, 1979; Lawson, 1979a, 1981a, b; Levson and Rutter, 1989a, b). Their relative thinness, local interbedding with sands and gravels, and common occurrence on the surface is also compatible with a supraglacial origin. Deposition probably occurred in the form of debris flows off the glacier surface or by gradual let-down by melting during deglaciation.

Unit 5 - Upper Gravel, Sand and Diamicton Sequences and Fine-Grained Sediments (Deglacial Deposits)

This unit consists of complex sequences of interbedded gravel, sand and / or diamicton deposits and local fine-grained sediments that invariably occur at or near the surface. Sequences containing diamicton are distinguished from Unit 4 by the relatively high proportion of sand and gravel strata in Unit 5 and by an erosional lower contact. Deposits of Unit 5 unconformably overlie unit 4 at all sites. Two subunits are recognized on the basis of the dominant sediment type.

Stratified Gravel, Sand and Diamicton Sequences (Proglacial Deposits)

Interbedded gravel, sand, diamicton and minor silt unconformably overlie Unit 4 at a number of locations including the Lower Portal Creek, Roche Bonhomme and Jacques Creek sites (Figs. 2-2b, 2-2c and 2-2d). Gravel and sand beds show horizontal stratification as well as planar and trough cross-stratification. Gravel beds are poorly to well sorted whereas sandy and silty strata are mainly well sorted. Clast lithologies are similar to the underlying glacial deposits (Fig. 2-4). Clasts are mainly moderately to well rounded and some are striated. Diamicton interbeds are commonly associated with these deposits but they are typically less abundant than sand and gravel and their relative proportion generally decreases towards the top of the subunit. Paleocurrent directions indicate valley-parallel flow. The lower contact of this subunit typically is erosional (Fig. 2-2).

Gravel, sand and diamicton strata of this subunit show highly variable dips and dip directions at some sites. Strata commonly dip steeply (up to 35°) towards the valley sides or rarely up valley. Normal faults and deformation structures indicating collapse towards the valley centre also occur in these deposits as do load and injection structures. Deformation structures generally are most abundant in the upper part of the subunit. Exposures of these deformed deposits are particularly common in terrace-like benches that flank the Athabasca River valley at various levels up to about 200 metres above the modern river (Figs. 2-8a and 2-8b). Sediments exposed in these terraces in the Jasper town site area are described in detail by Levson and Rutter (1989b). The terraces occur at different levels along the valley walls and they commonly are incised by large gullies with disproportionately small drainage areas.

Interpretation

Gravels and sands of this subunit are interpreted as ice-proximal glaciofluvial deposits. Their stratigraphic position indicates deposition during or shortly after deglaciation. At most locations, such as at the Lower Portal Creek and Jacques Creek sites (Figs. 2-2b and 2-2d), these interbedded gravels and sands show mainly down valley paleoflow directions and scoured lower contacts, suggesting that they were deposited as channel-fill sequences in proglacial streams issuing off the retreating glaciers (Levson and Rutter, 1986). Diamicton interbeds are similar to the glacially derived debris flow deposits in Unit 4 and their decrease in abundance towards the top of this unit probably reflects the decreased proximity of the glacial debris sources as deglaciation progressed.

Highly deformed deposits of this subunit showing paleoflow directions towards the valley sides are interpreted as proximal glaciofluvial and debris flow sediments deposited along the lateral margins of debris covered ice remaining in the valley centre.

The terrace-like benches that these deposits are associated with, in the Jasper town site area and elsewhere along the sides of the Athabasca valley, are interpreted as kame terraces. The presence of normal faults with large displacements and associated deformation structures indicate that substantial collapse occurred after deposition, probably as a result of melting of supporting ice in the valley centre. Boulton and Eyles (1979) and Bordonau (1993) described similar ice-marginal deposits along the flanks of former valley glaciers. The debris source for these features is believed to be ablating medial moraines and medial debris septa that extended into the Athabasca valley from tributary valleys (Levson and Rutter, 1989b). This interpretation is supported by the occurrence of the terraces below major stream confluences such as the Miette, Maligne, Snaring, Rocky and Snake Indian Rivers where medial moraines and debris septa would have been best developed. The ice-marginal kame terraces probably formed as debris washed and slumped down the margins of the ablating ice during the late stages of retreat onto the adjacent valley sides. Small lateral benches at different elevations along the valley walls may have formed as ice ablated to progressively lower levels. Large isolated gullies within the terraces, that are not connected to any significant modern stream drainage, may have formed by melting of localized ice cores within the terraces or by glacial meltwater incision shortly after their formation. These geomorphic characteristics support an ice-marginal or ice-contact origin for the deposits that comprise the terraces.

Silt and Clay Dominated Sequences (Glaciolacustrine Deposits)

At a few sites in the Athabasca River valley (*e.g.* Fig. 2-2g), diamicton beds of Unit 4 are overlain by massive to horizontally laminated silts and clays. These sediments are similar to those in Unit 2 and they are also interpreted as glaciolacustrine deposits. Their stratigraphic position indicates that they were deposited in ice-marginal lakes during retreat of glaciers occupying the Athabasca Valley. They are relatively rare compared to deposits in Unit 3, suggesting that during deglaciation, glaciers in the Athabasca valley did not dam drainage in tributary valleys as they did during the advance phase. It is therefore inferred that the Athabasca Valley became ice-free before or at about the same time as its tributary valleys. In the Punchbowl Falls area (Figs. 2-2g and 2-10), silts and clays of this subunit occur at the top of an incised bench on the southeast side of the valley and it is more likely that they were deposited in a lake confined there by ice than in an unconfined post glacial lake. These sediments contain various species of gastropods that yielded a radiocarbon date of 11,900 ± 120 years B.P. (GSC - 3885), giving a minimum date for retreat of ice from the region (Levson *et. al.,* 1989). The underlying gravels at this site are correlated with Units 1 and 2, rather than younger postglacial gravel units, because they are overlain by till of Unit 4 at nearby sites and they are lithologically and stratigraphically similar to Units 1 and 2 elsewhere in the area.. The absence of till of Unit 4 below the lake sediments at this site suggests that the till was not locally deposited or was eroded subglacially by ice or subglacial meltwater.

Unit 6 - Fine Sand, Silt and Diamicton Sequences (Post-Glacial Deposits)

Fine Sand and Silt Deposits (Loess Sequences)

Most sections in the region are capped by about 0.2 to 2 metres of massive fine sands and silts interpreted as loess (Figs. 2-2 and 2-11). These sediments are most common in the Pocahontas and Punchbowl Falls areas (Fig. 2-1b) where they overlie silt and clay dominated sequences in Unit 5. Fine sands and silts of Unit 6 locally contain buried volcanic ash and paleosol horizons. A prominent volcanic ash layer in silts overlying poorly sorted gravels at Pocahontas (Figs. 2-1b and 2-11b) has bracketing radiocarbon dates on charcoal of 4,180±210 years BP (AECV-602C) and 3,070±90 years BP (AECV-603C). The ash is interpreted as St. Helen's Yn tephra which was erupted about 3400 years BP (Mullineaux *et al.*, 1975; Westgate, 1980). Although not positively identified in the Pocahontas area, tephra chemically similar to Mount Mazama ash (Table 2-1), has been identified in reworked glaciolacustrine sediments elsewhere in the Athabasca River valley (Levson *et al.*, 1989). Intermittent deposition, and possible reworking, of volcanic ash occurred during deposition of the loess. Periods of non-deposition allowed for soil development. Glaciolacustrine silts of Unit 5, in combination with floodplain silts and sands in the Athabasca Valley, probably were the main source of the eolian deposits.

Diamicton Dominated Sequences (Paraglacial Fan Deposits)

Interbedded diamicton and poorly sorted gravels occur at the surface in a number of shallow exposures along the lower slopes of the Athabasca valley (Figs. 2-11a, 2-11b). These deposits are thickest at the base of steep valley slopes. They occur mainly at lower elevations in small fan-shaped landforms. Diamicton beds are mostly matrix supported, muddy and massive or crudely bedded. They contain mixtures of both angular, local clasts and glacially abraded, distally-derived clasts. Gravels are poorly sorted and may be matrix or clast supported. Interbedded sands and gravels occur mainly as thin trough-shaped lenses. Wood and charcoal fragments are common only in the uppermost parts of these deposits as are massive silt and fine sand beds containing multiple buried Bm and Ah soil horizons. Near Disaster Point (Fig. 2-1a), silt and fine sand deposits with several buried paleosols and numerous charcoal fragments are interbedded with diamicton and poorly sorted gravels of this subunit (Fig. 2-11a).

Interpretation

Diamicton beds in these deposits are believed to have originated as valley-side, debris flows that incorporated previously deposited glacial sediments as well as local bedrock materials. Resedimentation of debris along the valley walls probably was most active during and shortly after deglaciation when environmental conditions were most conducive to slope instabilities (Ryder, 1971; Jackson *et al.*, 1982). The occurrence of wood only in the upper parts of these deposits suggests a restricted forest cover at least during their initial depositional phase. The occurrence of the resedimented materials in low elevation fans suggests that they were deposited in early post-glacial times after some stream incision had taken place. Sand and gravel lenses interbedded with diamicton beds in this subunit probably were deposited in small rills and fan channels by surface runoff. Massive silt and fine sand beds that occur in the upper one or two metres of the diamicton and gravel sequences are interpreted as loess deposits.

CHRONOSTRATIGRAPHY

The informal lithostratigraphic units described above are placed here into a general chronostratigraphic framework for the region, mainly on the basis of radiocarbon dates and some tephrochronologic control. Materials suitable for radiocarbon dating from the paleofan sediments of Unit 1 are rare but wood fragments (Picea *sp.*), obtained from an exposure at Amber Mountain (Fig. 2-3a), were dated at >39,750 (AECV 450c) and >48,000 (GSC-4646HP). Since the wood occurs at a relatively high level in the section, a large part of the fan sequence at that site must have accumulated prior to 48 ka. Deposits in the upper part of the paleofan sequences are locally interbedded with overlying glaciofluvial gravels indicating that alluvial fan sedimentation must have continued until the onset of the last glaciation. In addition, as ice did not reach the area until sometime after 29 ka (see below), fan sedimentation must have continued for at least an additional 20,000 years after deposition of the wood bearing horizon. Thus, sedimentation in the Amber Mountain fan probably spanned most of the Middle Wisconsinan and probably part of the Early Wisconsinan. The age of the oldest fan deposits in Unit 1 is unknown but the presence of rare glacial erratics throughout the sequences indicates that they began to accumulate sometime after the first Pleistocene glaciers reached the area. Evidence of earlier Pleistocene glaciations in the Jasper region is restricted mainly to these erratic clasts and to some high level erratics that occur at elevations above the locally inferred level of the last glaciation (see Chapter 5). The poor lithification and cementation of the paleofan deposits is also consistent with a Pleistocene age.

Although the glacial and associated glaciofluvial sediments in Units 2 to 5 comprise a group of sedimentologically complex deposits, they are interpreted to be the product of only one glacial event. The age of this event is constrained by a radiocarbon date of 29,100 ± 560 years B.P. (GSC 3792) obtained on wood in gravels of Unit 2 underlying till near Jasper town site. This date provides a maximum age for the onset of the last glaciation and indicates that ice moved into the area during the Late Wisconsinan stadial. A minimum date for deglaciation is provided by gastropods from ice-marginal lake sediments in the Pocahontas area, dated at 11,900 ± 120 years B.P. (GSC 3885). Numerous other limiting radiocarbon dates from at least four different high elevation areas in Jasper Park, ranging from 9445 ± 375 years B.P. (Beta 1480) to 9660 ± 280 years B.P. (BGS 465), indicate that glaciers throughout the region had receded to near their present limits by shortly after 10 ka (Luckman and Osborn, 1979; Kearney and Luckman, 1983a, b; Beaudoin, 1984; Luckman and Kearney, 1986; Osborn and Luckman, 1988). Only minor advances, extending about one to two kilometers beyond present glacier margins, have since occurred (Kearney and Luckman, 1981; Osborn and Luckman, 1988).

Paraglacial conditions at the end of the last glaciation resulted in widespread resedimentation of glacigenic debris in the region and the development of numerous paraglacial fans (see Unit 6) in the Athabasca valley. Radiocarbon dates on charcoal in the loess deposits blanketing fan surfaces near Pocahontas indicate that most fan sedimentation had ceased by about 7500 years ago (Levson et al., 1989). A number of dates ranging from approximately 4000 yrs BP to 7250 yrs BP have been obtained from stratigraphically equivalent loess deposits in the region (Dumanski et al., 1980; Levson et al., 1989). St. Helen's Yn tephra occurs within the loess sequence in the Pocahontas area and volcanic ash chemically similar to Mount Mazama tephra (Table 2-1), dated at about 6800 years BP (Bacon, 1983), occurs near the surface in alluvial fan deposits reworked from glaciolacustrine silts near Athabasca Falls. Similar stratigraphic relationships between Mount Mazama ash and paraglacial alluvial fan deposits were described in central British Columbia by Ryder (1971) who concluded that the bulk of fan sedimentation occurred shortly after deglaciation when paraglacial conditions resulted in unstable slopes and high depositional rates. This environment probably existed for a relatively short period of time following deglaciation. Roed and Wasylyk (1973) and Jackson et al. (1982) also concluded that post-glacial alluvial fan sequences in the southern Canadian Rocky Mountains were deposited mainly in paraglacial conditions in the Late Pleistocene. Most of the low elevation alluvial fan sequences of Unit 6 in the Athabasca River valley, therefore were probably deposited during the Late Pleistocene and/or early Holocene.

SUMMARY

The Athabasca River in Jasper National Park occupies a broad, low-relief valley within an overall high-relief region in the Rocky Mountains. The valley acted as a depositional sink during the Late Pleistocene and allowed for the accumulation of a locally thick sequence of Quaternary deposits. Exposures of these sediments are dominated by thick accumulations of paleoalluvial fan deposits as well as a complex glacial sequence associated with the last glaciation. The paleoalluvial fan deposits (Unit 1) are characterized by coarse-grained, gravelly diamicton beds, interpreted as debris flow and flood flow deposits, with interstratified gravels and minor sands deposited by intermittent streams. Radiocarbon dates on wood in the fan deposits indicate that the fans began to accumulate well before 48 ka and probably continued to develop until Late Wisconsinan glaciers advanced into the area. Deposits resulting from fluvial sedimentation in the valley bottoms during this non-glacial period either were eroded or buried in most areas, but braided stream gravels preserved near Jasper town site were deposited about 29 ka. These gravels (Unit 2, lower subunit) are clast-supported, well imbricated and horizontally bedded, and are interpreted as longitudinal bar deposits. Wood and shell fragments occur throughout this subunit but decrease in abundance towards the top.

The transition from alluvial fan sedimentation in Unit 1 to braided stream deposition in Unit 2, reflects the increasing influence of outwash systems that developed in front of glaciers advancing down the main valleys. Local interbedding of diamicton beds, of probable alluvial fan origin, and a gradual rather than sudden increase in abundance of distally derived rocks (between Units 1 and 2), suggests that glaciofluvial sedimentation was locally coeval with the latter stages of alluvial fan development. The high elevation of the glaciofluvial gravels at most sites compared to present day rivers indicates a higher local base level, probably reflecting aggradation in the main valleys prior to glaciation. An increase in the number of cut and fill structures, erratic clasts and coarse lag gravels as well as a decrease in abundance of organics in the upper part of Unit 2, are inferred to reflect a more proximal glaciofluvial origin for these gravels than for those in the lower subunit. Large variations in sediment sorting, grain-size distribution, clast size and shape, and paleocurrent direction in the upper subunit, as well as the presence of glacially-derived debris flow deposits, also indicate a proximal ice-marginal environment.

Parallel laminated silts and clays (Unit 3) overlying glaciofluvial gravels in Unit 2 in some tributary valleys of the Athabasca River valley, particularly in the Front Ranges, are interpreted as glaciolacustrine sediments deposited in ice-dammed lakes. These sediments locally coarsen upwards and contain interbedded subaqueous debris flow deposits in their upper part. This is believed to reflect the progressive deposition of more proximal glaciolacustrine deposits as glaciers advanced down these valleys.

Massive, matrix-supported, compact, diamicton sequences (Unit 4), unconformably overlying deposits of Units 1 to 3, are interpreted as basal tills deposited by the advancing glaciers. These deposits exhibit strong valley-parallel fabrics and they contain numerous glacially abraded and erratic clasts. Lodged clasts and glacially-induced folds and thrust faults occur in sediments directly underlying this unit at some sections. Massive diamictons in Unit 4 may be underlain or overlain by stratified, sandy diamicton, commonly with poorly developed fabrics and interbedded sand and gravel lenses. These deposits are interpreted as glacially derived, debris flow and glaciofluvial sediments. They were deposited during both the advance and retreat stages of glaciation. Angular metamorphic erratics derived from west of the Continental Divide, locally occur in the Athabasca River valley, mainly on the surface in association with a sandy, poorly consolidated, diamicton. They are inferred to be part of the Athabasca Valley Erratics Train and the associated diamicton is interpreted as a supraglacial till.

During deglaciation, a complex sequence of deposits (Unit 5) consisting of interbedded gravel, sand and diamicton as well as local clay and silt sediments was deposited on the surface at most sites. Gravel and sand-dominated strata exhibit characteristics of proximal glaciofluvial deposits and interbedded diamictons are interpreted as glacigenic debris flow deposits. These sediments are preserved mainly as proglacial channel-fill gravels and as kame terrace sequences deposited along the flanks of the Athabasca River valley. Kame terrace deposits show large variations in paleoflow directions and abundant post-depositional deformation structures including numerous normal faults indicating collapse towards the valley centre. Silt and clay dominated sequences in Unit 5 are relatively rare, occurring at only a few sites where small ice-marginal lakes formed during deglaciation. The absence of glaciolacustrine sediments at the surface, in most valleys where advance-phase glacial lakes formed, suggests that the Athabasca River valley probably became ice-free before or at approximately the same time as its tributary valleys. Radiocarbon dating of shells in glaciolacustrine silts at Pocahontas suggest that deglaciation occurred there sometime after 12 ka.

Shortly after deglaciation, much of the unconsolidated glacial debris flanking the valley sides was resedimented under paraglacial conditions resulting in deposition of numerous fan sequences at the base of the valley slopes. Deposition occurred mainly in early post-glacial times, as indicated by radiocarbon dates on wood and by the presence of Mazama ash (6800 years BP) near the top of the fan deposits. Reworking of older deposits, especially sandy glaciolacustrine and glaciofluvial sediments, by wind, also occurred in late-glacial and post-glacial times and resulted in widespread deposition of a loess mantle over the region.

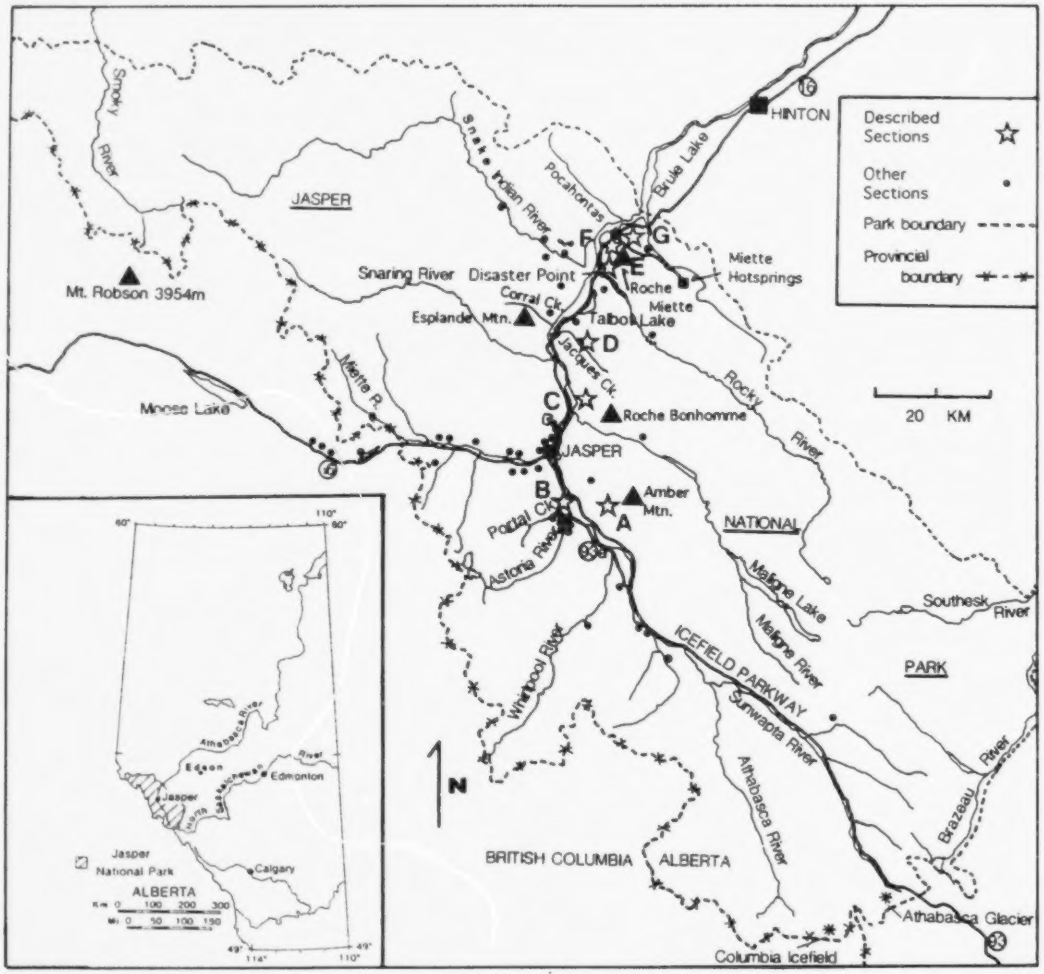

HINTON
Described Sections ☆
Other Sections ●
Park boundary
Provincial boundary
Smoky River
JASPER
Snake Indian River
Pocahontas
Brule Lake
F
G
E
Miette Hotsprings
Roche Miette
Mt. Robson 3954m
Snaring River
Disaster Point
Corral Ck.
Esplande Mtn.
Talbot Lake
D
Jacques Ck.
Rocky River
20 KM
Moose Lake
Miette R.
C
Roche Bonhomme
JASPER
B
Amber Mtn.
Portal Ck.
A
Astoria River
NATIONAL
Maligne Lake
Maligne River
Southesk River
ICEFIELD PARKWAY
Whirlpool River
Sunwapta River
PARK
N
Athabasca River
Brazeau River
BRITISH COLUMBIA
ALBERTA
Athabasca Glacier
Columbia Icefield
Athabasca River
Edson
Edmonton
Jasper
North Saskatchewan
River
Calgary
Jasper National Park
ALBERTA
60°
110°
49°
114°

Fig. 2-1a. Location map of measured lithostratigraphic sections and place names in the Jasper National Park region. Sites labeled A to G correspond to sections shown in Figure 2-2.

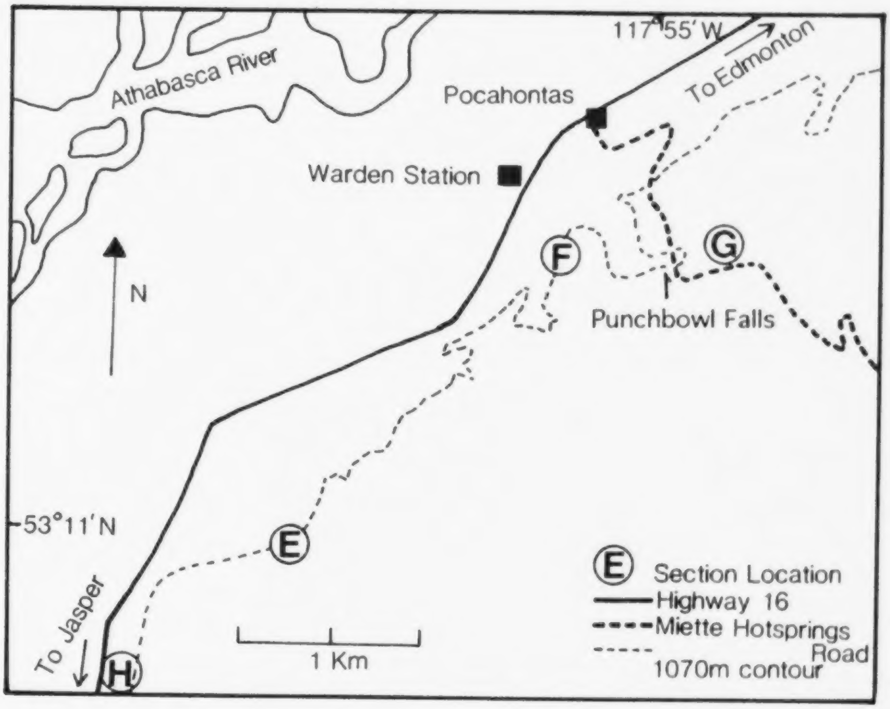

117°55′W
To Edmonton
Athabasca River
Pocahontas
Warden Station
N
F
G
Punchbowl Falls
53°11′N
E
To Jasper
H
1 Km
E Section Location
Highway 16
Miette Hotsprings Road
1070m contour

Fig. 2-1b. Detailed location map of sections discussed in the text in the Pocahontas area. Sites labeled E to G correspond to sections shown in Figure 2-2. Site H is the location of the section provided in Fig. 2-11a.

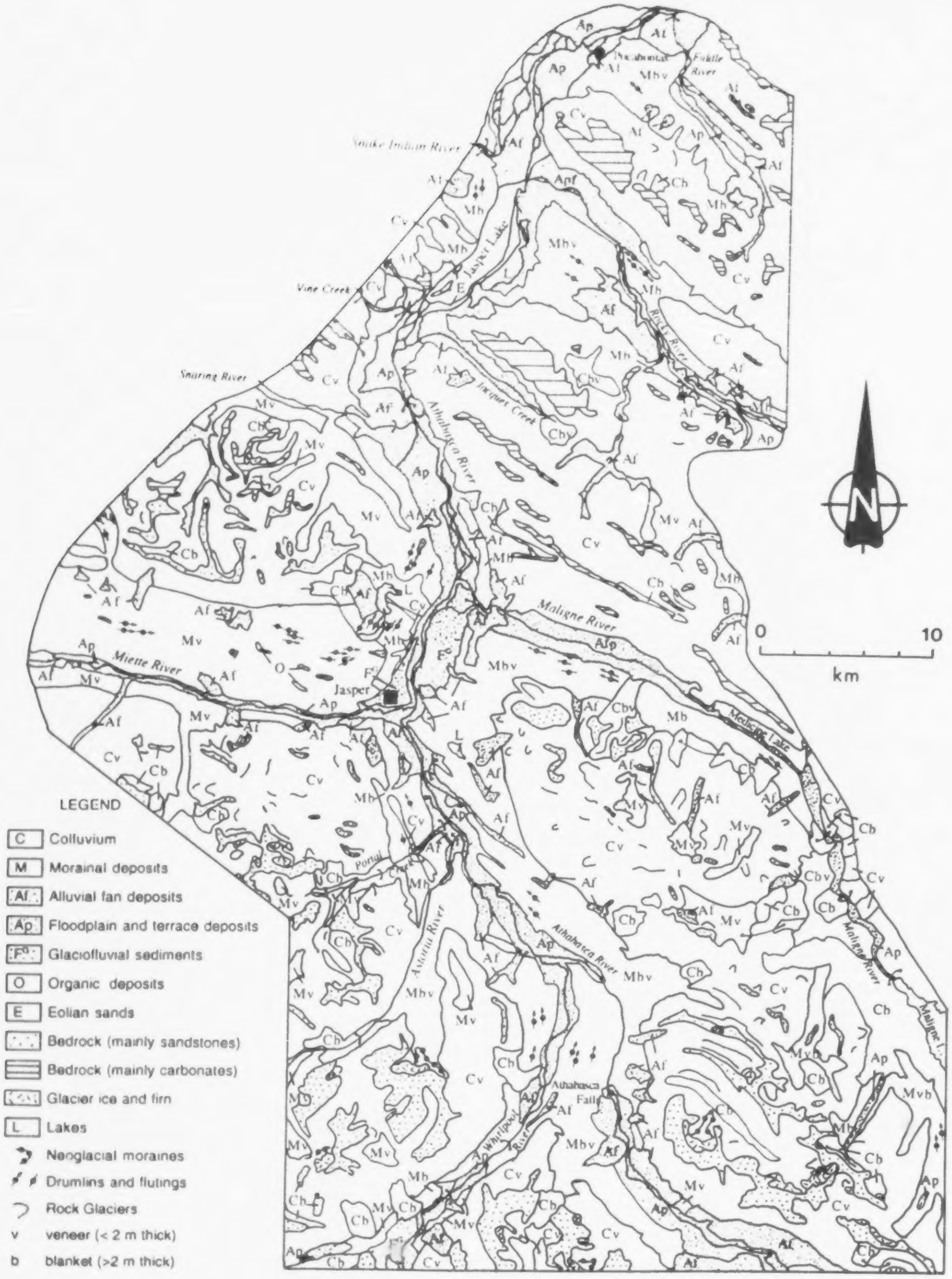

LEGEND
C Colluvium
M Morainal deposits
Af Alluvial fan deposits
Ap Floodplain and terrace deposits
F^G Glaciofluvial sediments
O Organic deposits
E Eolian sands
Bedrock (mainly sandstones)
Bedrock (mainly carbonates)
Glacier ice and firn
L Lakes
Neoglacial moraines
Drumlins and flutings
Rock Glaciers
v veneer (< 2 m thick)
b blanket (>2 m thick)
N
0
10
km
Pocahontas
Fiddle River
Snake Indian River
Jasper Lake
Vine Creek
Snaring River
Rocky River
Jacques Creek
Athabasca River
Maligne River
Miette River
Jasper
Medicine Lake
Portal Creek
Astoria River
Whirlpool River
Athabasca Falls
Maligne River

Fig. 2-1c. Surficial geology map of the study area showing the present distribution of Late Wisconsinan morainal deposits (M), Neoglacial moraines and other surficial sediments (Compiled and modified from unpublished data by Reimchen [1976] and [Holland and Coen, 1982]).

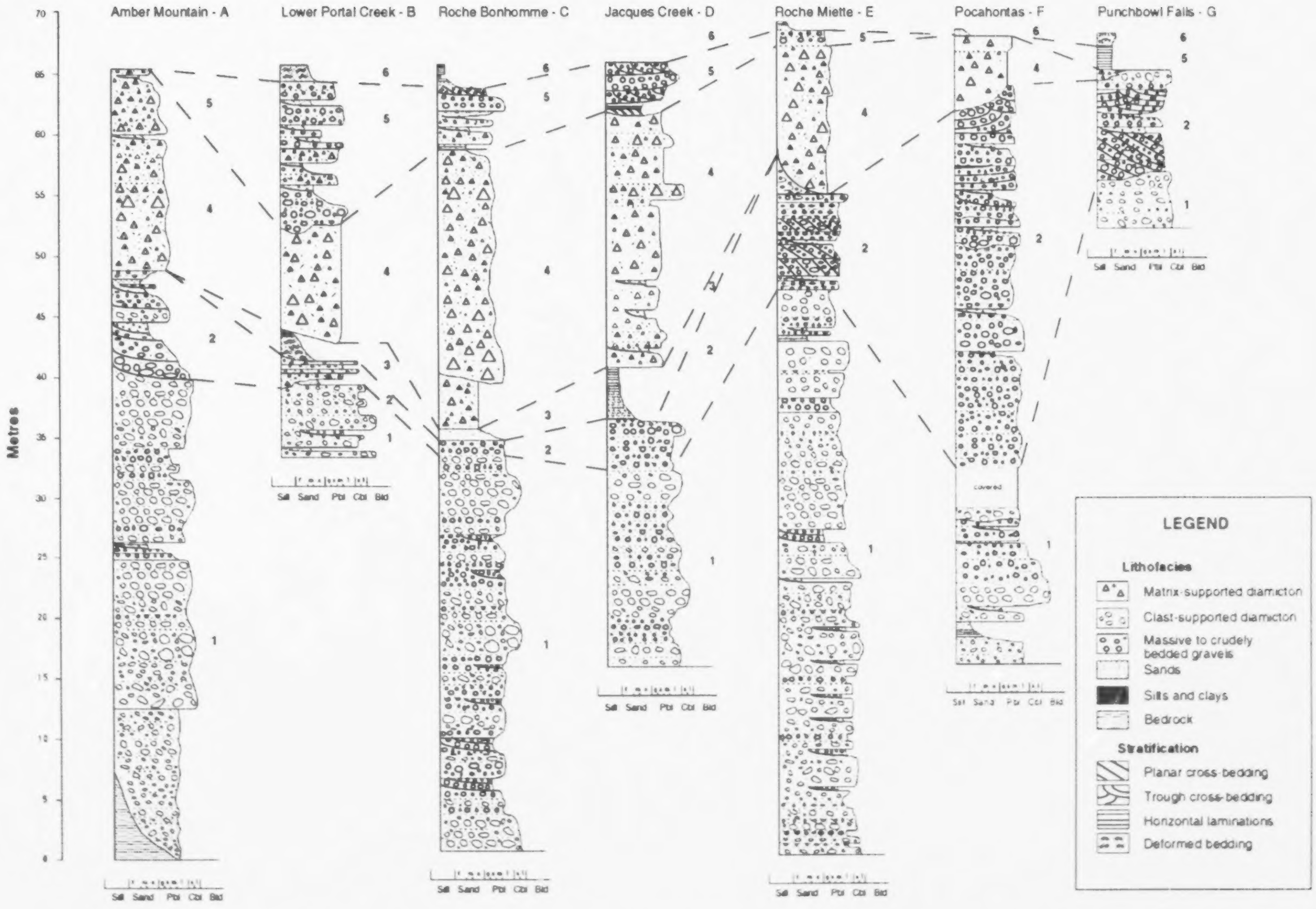

Amber Mountain - A
Lower Portal Creek - B
Roche Bonhomme - C
Jacques Creek - D
Roche Miette - E
Pocahontas - F
Punchbowl Falls - G
Metres
70
65
60
55
50
45
40
35
30
25
20
15
10
5
0
1
2
3
4
5
6
Covered
Silt Sand Pbl Cbl Bld
LEGEND
Lithofacies
Matrix-supported diamicton
Clast-supported diamicton
Massive to crudely bedded gravels
Sands
Silts and clays
Bedrock
Stratification
Planar cross-bedding
Trough cross-bedding
Horizontal laminations
Deformed bedding

Fig. 2-2. Composite stratigraphic columns from each of the main study sites and units discussed in the text. Section locations are shown on Figs. 2-1a and 2-1b. Unit 1 - paleofan deposits; Unit 2 - glaciofluvial gravels and sands and minor proglacial debris flows; Unit 3 - glaciolacustrine silts and clays; Unit 4 - till and associated glacigenic debris flows; Unit 5 - late-glacial glaciolacustrine and glaciofluvial sediments, paraglacial debris flow deposits and minor loess.

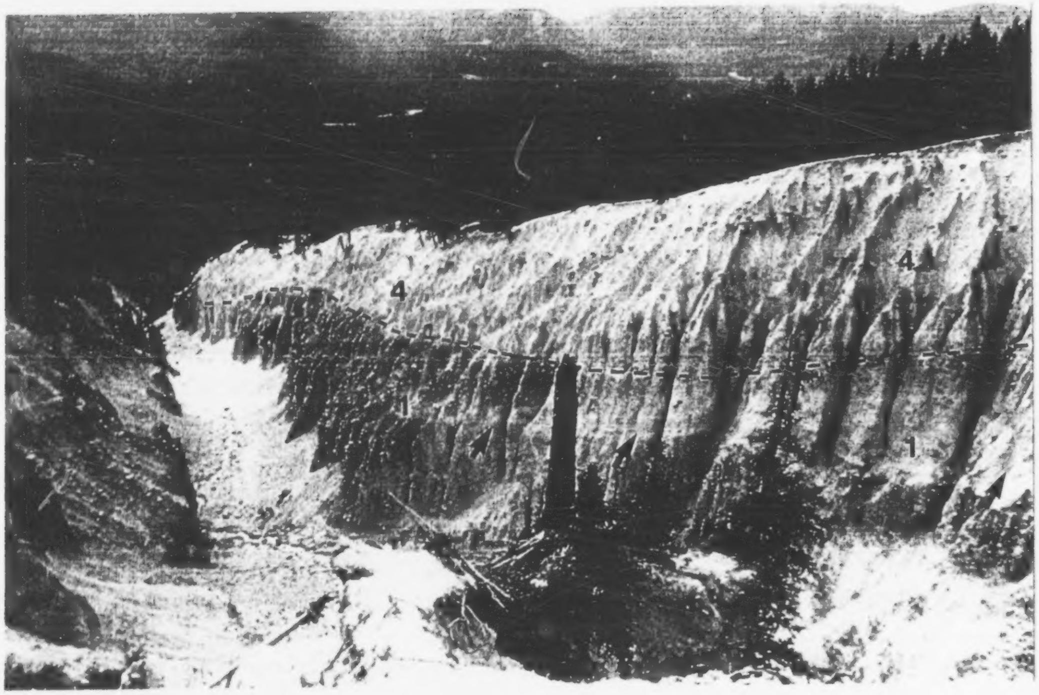


Fig. 2-3. a) Paleoalluvial fan, debris-flow deposits (Unit 1) unconformably overlain by matrix-supported, glacial deposits (Unit 4) at Amber Mountain (location A on Fig. 2-1a; see also Fig. 2-2a). Note the crude bedding dipping down valley, coarse grain-size and overall poor sorting of the alluvial fan deposits. Wood fragments yielding infinite radiocarbon dates were recovered from the prominent, light coloured, sand and gravel horizon (arrowed) in the upper part of the fan deposits (view looking down fan to the northwest; Athabasca River valley in distance).

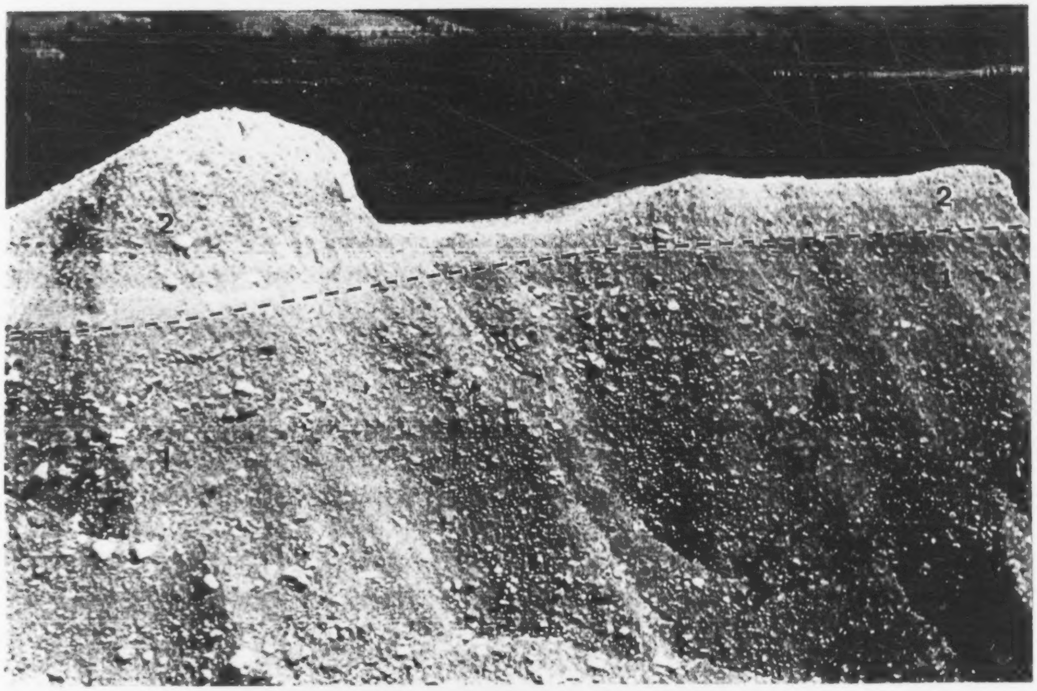


Fig. 2-3. b) angular, poorly sorted alluvial fan debris flow deposits (Unit 1) at the Roche Miette section (location E on Figs. 2-1a and 2-1b) overlain by glaciofluvial gravels of Unit 2 (light coloured unit at top).

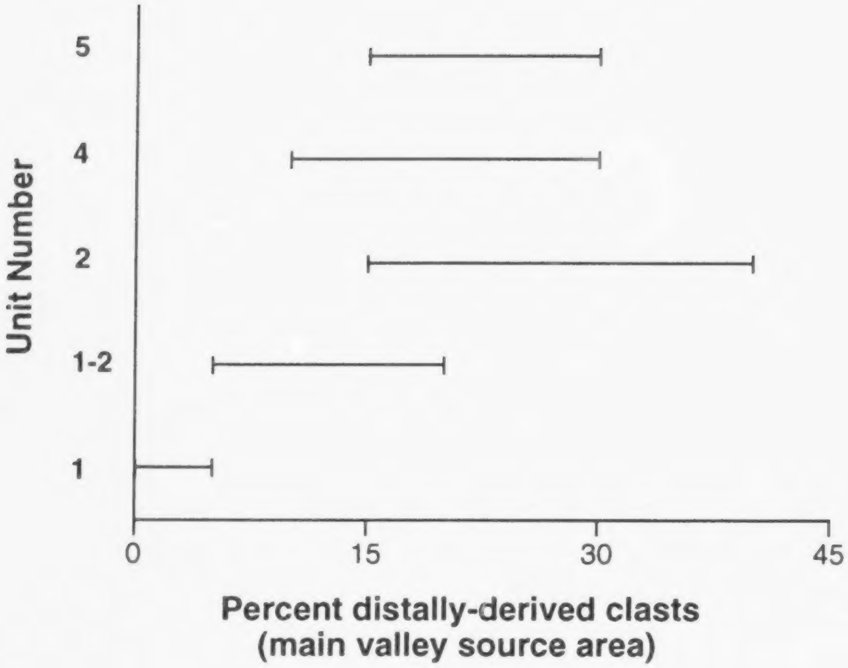


Fig. 2-4. Ranges in percentage of distally-derived clasts in main stratigraphic units in the study area. 'Distal' clasts include those derived from within the main valley drainages as well as far-traveled erratics (*e.g.* metamorphics) but excludes local tributary basin rocks.

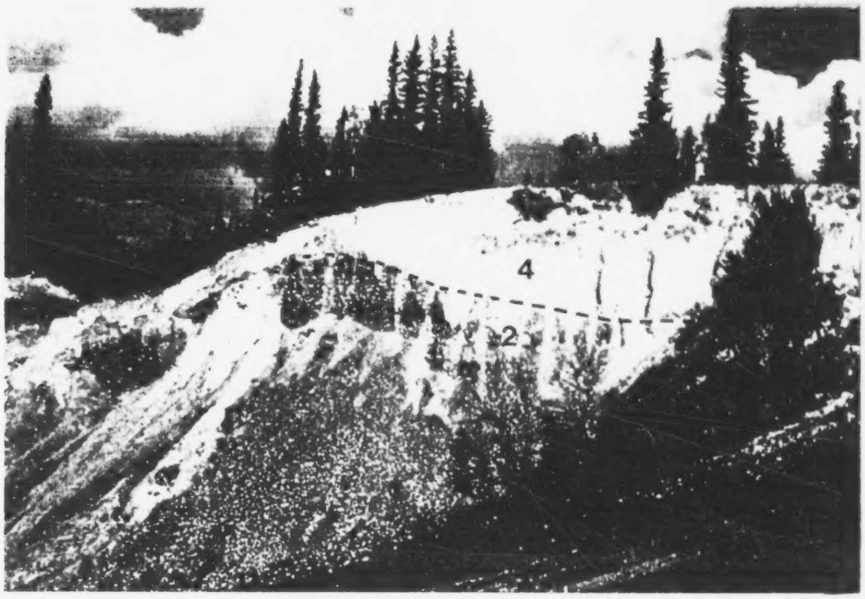


Fig. 2-5. a) Till of Unit 4 unconformably overlying glaciofluvial gravels (Unit 2) at the Pocahontas section (location F on Figs. 2-1a and 2-1b; see also Fig. 2-2f).

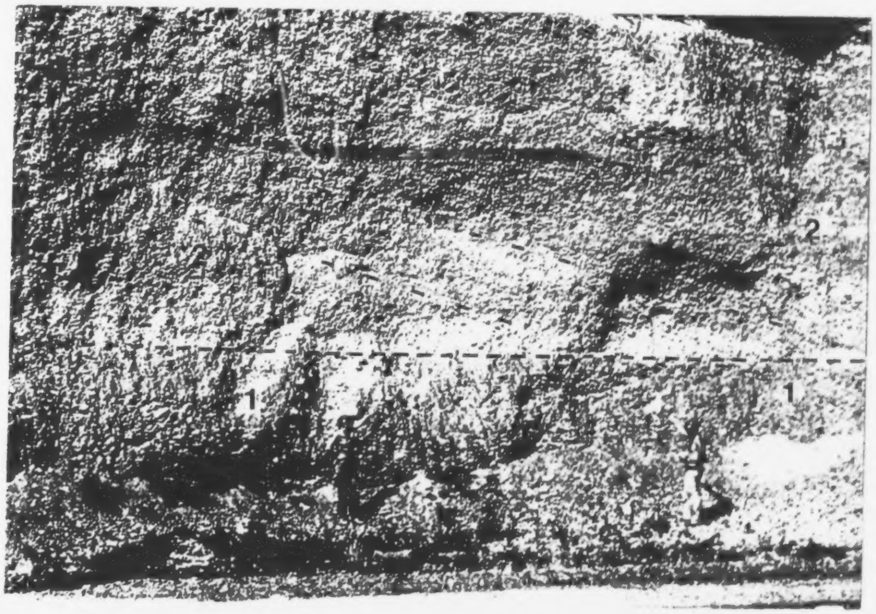


Fig. 2-5. b) glaciofluvial gravels (Unit 2) overlying debris flow deposits (base) of Unit 1 at the Punchbowl Falls section (location G on Figs. 2-1a and 1b; Fig. 2-2g). Note cross-bedding in lower part of Unit 2.

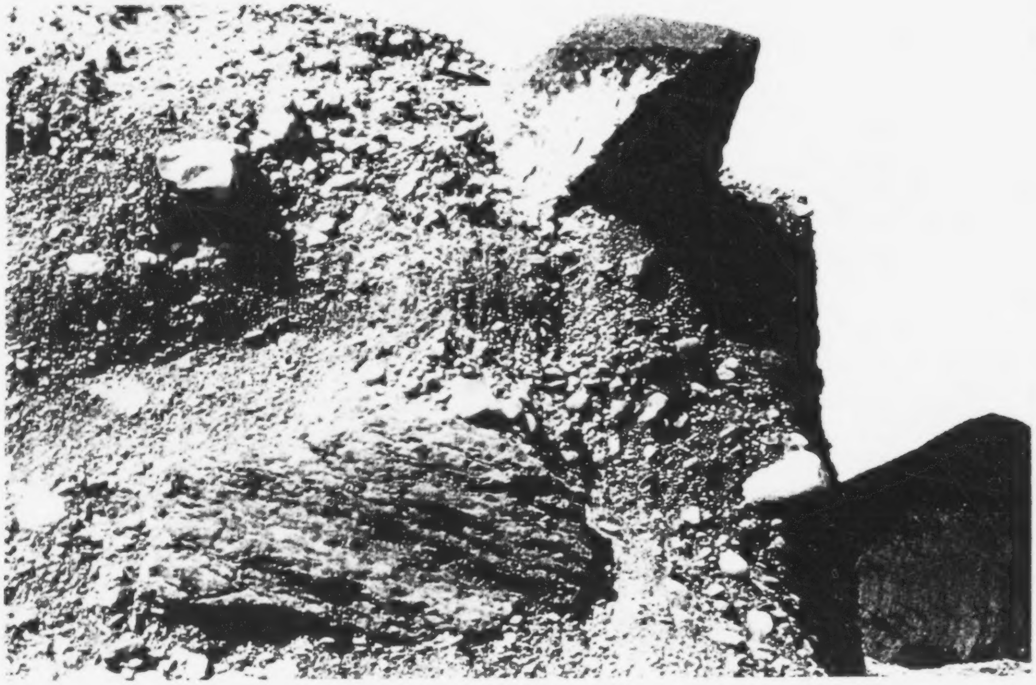


Fig. 2-5. c) deformed gravels and lodged boulder (arrowed at top) in the upper part of the glaciofluvial gravels shown in 3b); the upper surface of the lodged boulder is faceted and striated, with striae trending parallel to the main valley. Note differences in clast size, sorting and deformation in comparison to undeformed gravels in Figs. 2-3b and 2-5b.

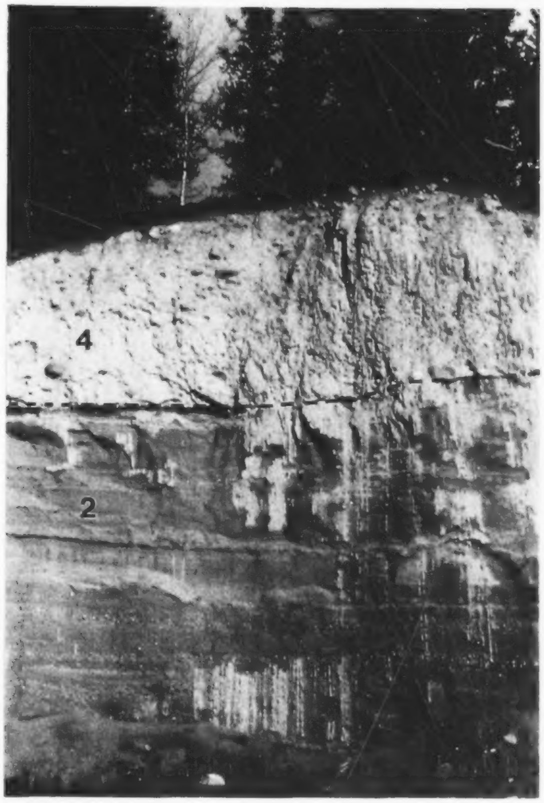


Fig. 2-6. a) Late Wisconsinan till (Unit 4) erosionally overlying Middle Wisconsinan, braided stream gravels near Jasper townsite (Unit 2). Radiocarbon dating of wood recovered from the upper most sand and gravel horizon shown indicates that deposition of this unit occurred about 29,100 ± 560 years B.P. (GSC-3792).

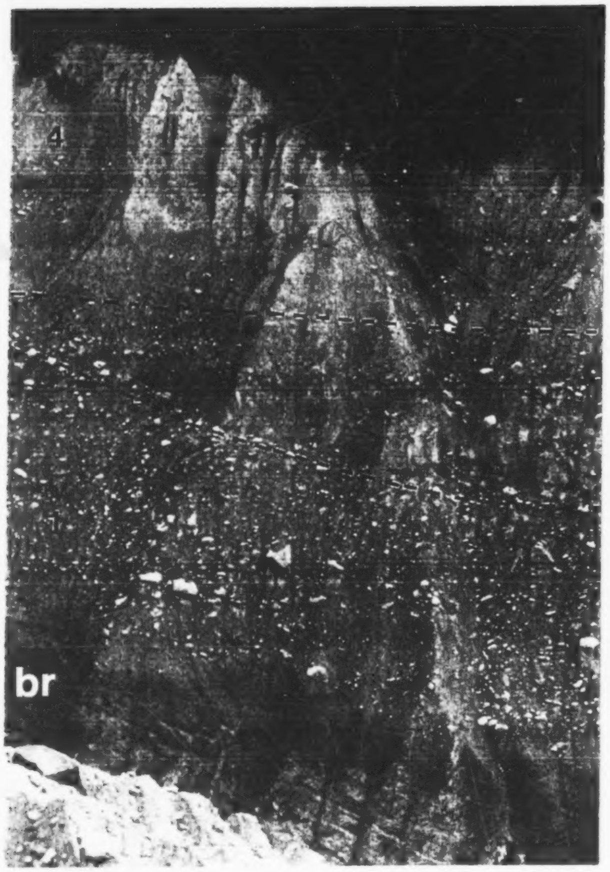


Fig. 2-6. b) matrix-supported till of Unit 4 (top) unconformably overlying channelized glaciofluvial gravels and sands of Unit 2 (centre) and clast-supported, bouldery, alluvial fan deposits in Unit 1 (bottom) at Amber Mountain (location A on Fig. 2-1a). Dark unit at base of section (br) is bedrock.

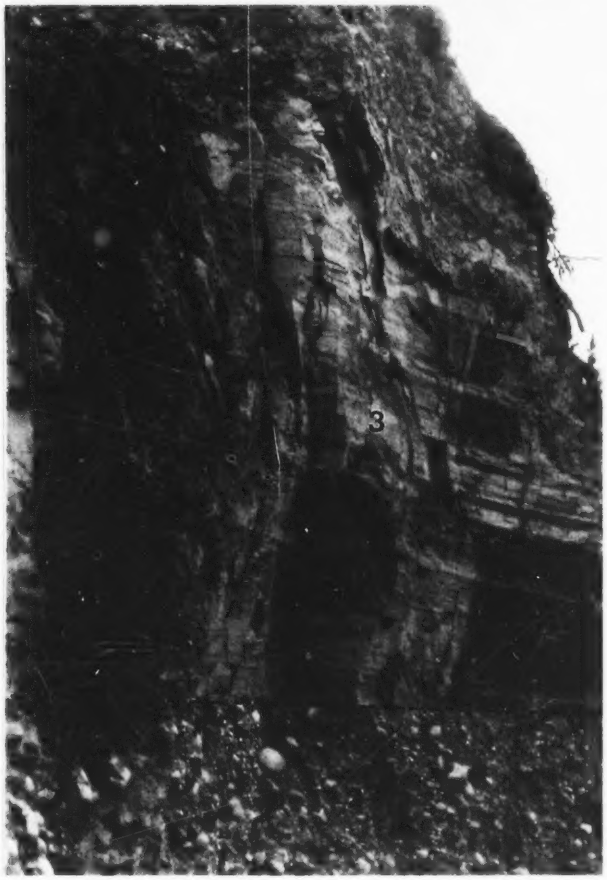


Fig. 2-7. Glaciolacustrine deposits of Unit 3 exposed along the lower reaches of the Snake Indian River valley near its confluence with the Athabasca River: a) horizontally stratified, glaciolacustrine silts and clays (distal, lake basin deposits) at the base of Unit 3, sharply but conformably overlying glaciofluvial sands and gravels of Unit 2 (silt and clay exposure is 4.25 m thick).

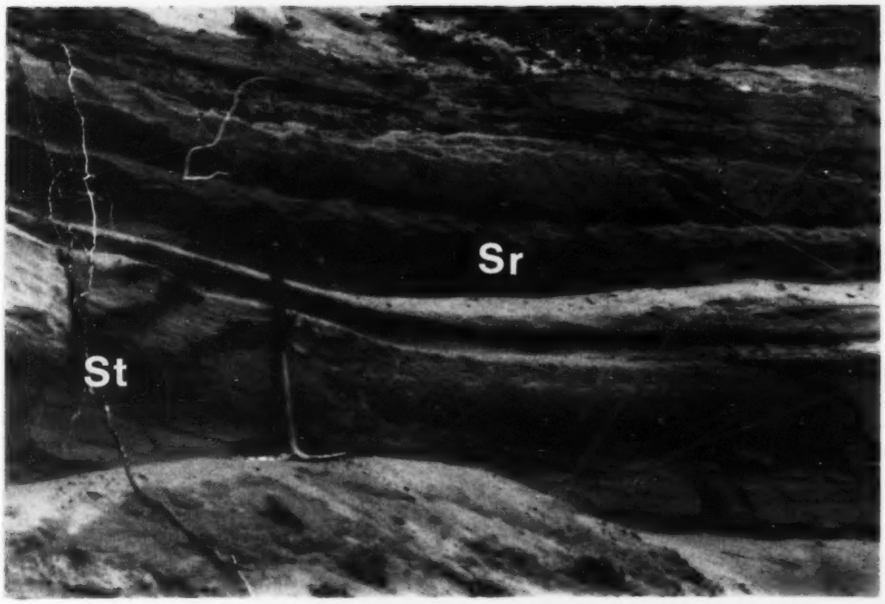


Fig. 2-7. b) climbing ripples (Sr) and trough cross-bedded sands (St) in the upper part of Unit 3, interpreted as proximal glaciolacustrine deposits.

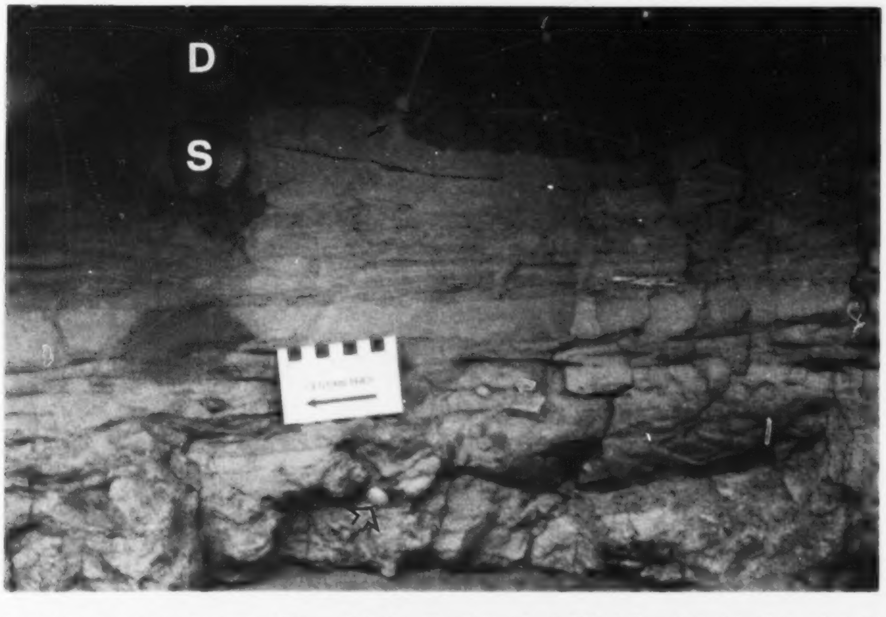


Fig. 2-7. c) deformed sandy strata (S) overlain by diamicton bed (D), inferred to be of a subaqueous debris flow origin. Note the load and injection structures (arrowed) at the base of the diamicton and dropstone below scale (open arrow).

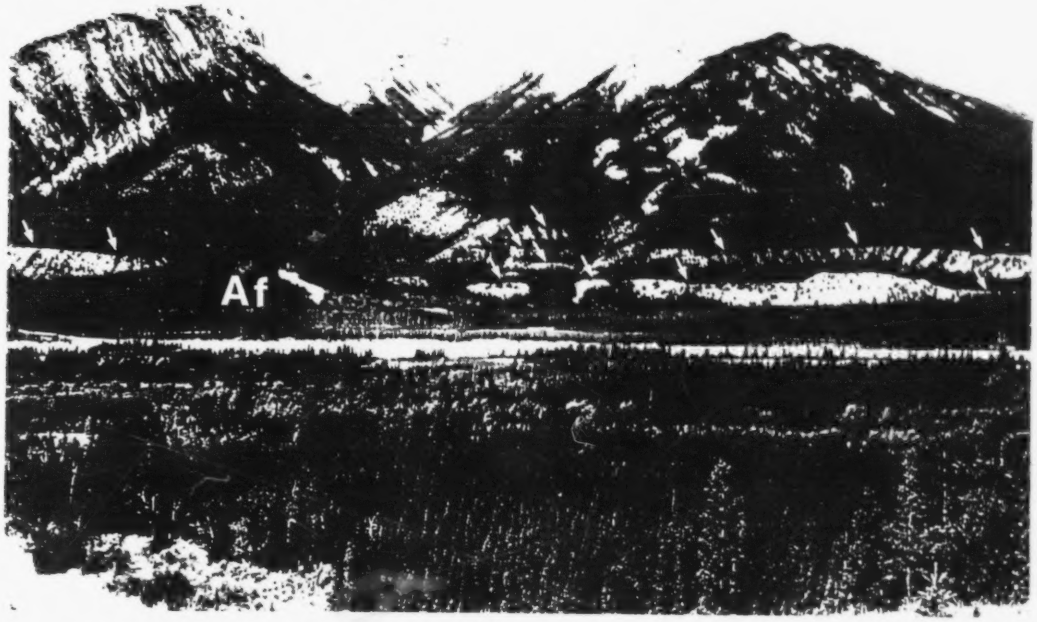


Fig. 2-8. a) Glacial and glaciofluvial terrace-like benches (arrowed) flanking the west side of the Athabasca River valley below Esplanade Mountain, opposite the mouth of the Jacques Creek (Fig. 2-1a); Note the variable slopes and levels of the benches and Corral Creek alluvial fan (Af) at centre-left.

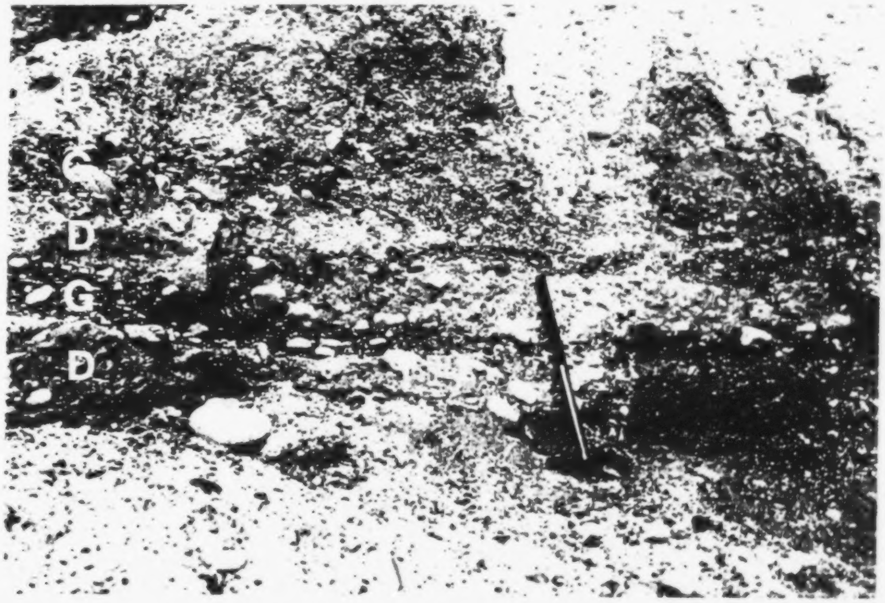


Fig. 2-8. b) Typical exposure of interbedded debris flow deposits (D) and fluvial gravels (G) in Unit 5.

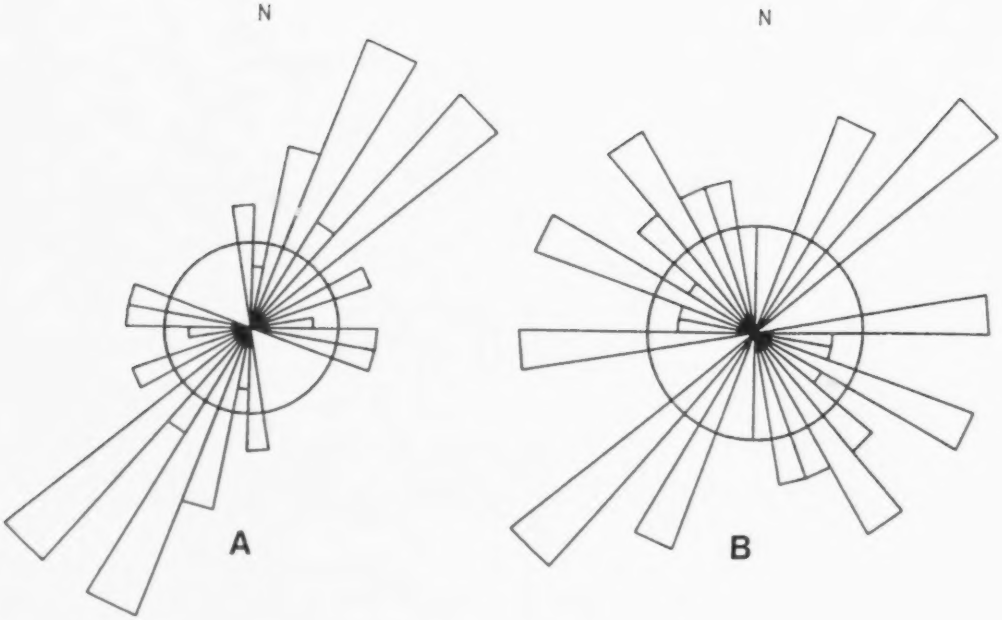


Fig. 2-9. Representative rose diagrams of a-axis orientations in matrix-supported diamicton units interpreted as a) till (S1= 0.74, N= 25) and b) debris flow deposits (S1=0.55, N=25). Fabrics were measured in Unit 4 at the Pocahontas section (Fig. 2-2f, location E on Fig. 2-1b). The main valley orientation at this site is northeasterly.

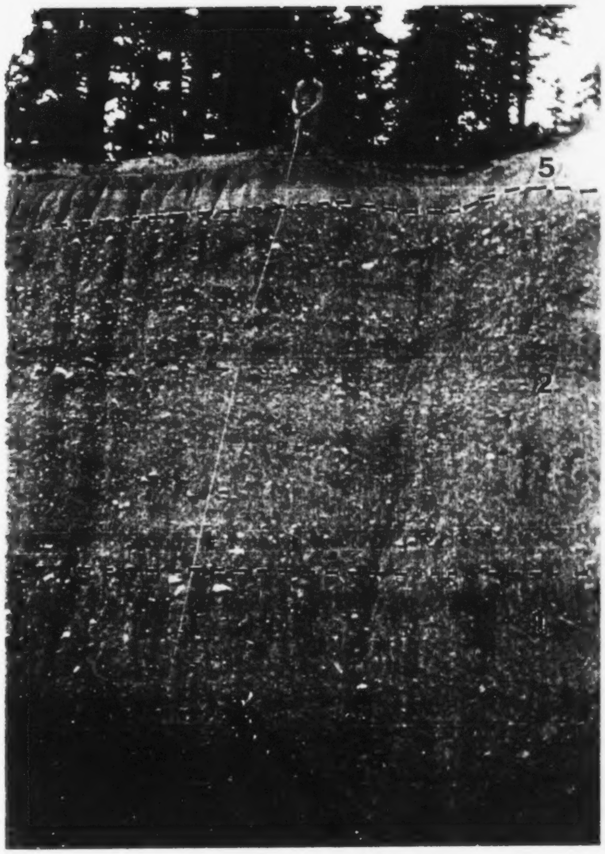
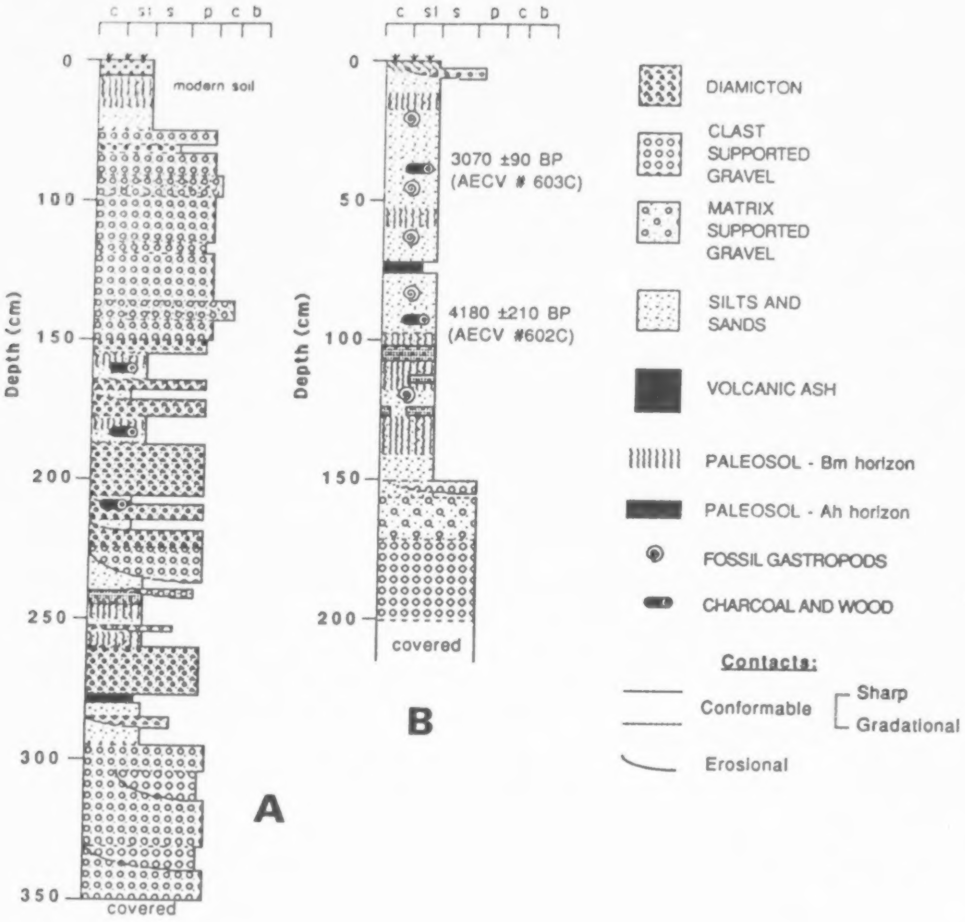


Fig. 2-10. Ice-marginal, glaciolacustrine silts and clays of Unit 5 unconformably overlying gravels of Units 1 and 2 in the Pocahontas area (location G on Fig. 2-1b). A radiocarbon date of 11,900 ± 120 years B.P. (GSC - 3885) on gastropod shells within the silts and clays provides a minimum date for retreat of Late Wisconsinan glaciers in the Athabasca River valley. Measuring tape is extended 15 metres.



c si s p c b
0
modern soil
100
150
200
250
300
350
Depth (cm)
covered
A
c si s p c b
0
50
100
150
200
3070 ±90 BP (AECV # 603C)
4180 ±210 BP (AECV #602C)
Depth (cm)
covered
B
DIAMICTON
CLAST SUPPORTED GRAVEL
MATRIX SUPPORTED GRAVEL
SILTS AND SANDS
VOLCANIC ASH
PALEOSOL - Bm horizon
PALEOSOL - Ah horizon
FOSSIL GASTROPODS
CHARCOAL AND WOOD
Contacts:
Conformable
Sharp
Gradational
Erosional

Fig. 2-11. a) Stratigraphic section of the upper few metres of a small alluvial fan near Disaster Point (location H on Fig. 2-1b). Paleosol horizons indicate periods of temporary stability on the fan surface between deposition of debris flow deposits (diamicton) and alluvial gravels;

b) Radiocarbon dates, volcanic ash and paleosol horizons in loess deposits unconformably overlying glaciofluvial gravels in the Pocahontas area (location F on Fig. 2-1b).

Sample #	841	842	843	844	Average	Regional Average*
Si	32.72	33.93	33.57	32.26	33.12 ±.0.66	33.11 ±0.29
Ti	0.27	0.30	0.26	0.33	0.29 ±0.03	0.29 ±0.01
Al	7.43	7.41	7.47	7.23	7.39 ±0.09	7.70 ±0.08
Fe	1.54	1.56	1.60	1.60	1.57 ±0.03	1.54 ±0.04
Ca	1.05	0.94	0.98	0.95	0.98 ±0.04	1.13 ±0.03
Na	3.59	3.53	3.72	3.49	3.58 ±0.09	3.80 ±0.11
K	2.33	2.44	2.37	2.44	2.39 ±0.05	2.17 ±0.03
Cl	0.27	0.03	0.26	0.28	0.21 ±0.10	0.18 ±0.02

* average of 14 samples from northwest USA and southwest Canada (Smith and Westgate, 1969)

Table 2-1. Chemical composition of glass shards from tephra interpreted as Mount Mazama ash in alluvial fan deposits near the confluence of the Whirlpool and Athabasca Rivers.

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CHAPTER 3 - SEDIMENTATION IN PRE-LATE WISCONSINAN ALLUVIAL FANS IN THE CANADIAN ROCKY MOUNTAINS: LITHOFACIES CHARACTERISTICS, MODERN ANALOGUES AND BEDROCK CONTROLS

INTRODUCTION

Thick successions of alluvial fan sediments underlying Late Wisconsinan glacial deposits are well exposed at several localities in the Rocky Mountains of Jasper National Park, Alberta, Canada (Figs. 3-1 and 3-2). These sediments are dominated by clast-supported, mud-poor, gravelly lithofacies that are both poorly sorted and crudely bedded. The lack of mud-rich cohesive debris flow deposits in these successions makes them distinct from arid-region fans, from which many models of alluvial fan sedimentation have been derived (*e.g.* Bull, 1972; Collinson, 1986; Rust and Koster, 1984). The low proportion of mud in the deposits also distinguishes them from late Pleistocene and early Holocene periglacial (*e.g.* Wasson, 1977) and paraglacial (Ryder, 1971; Eyles and Koscis, 1988) fan deposits and from modern alluvial fan successions formed in humid temperate regions (Kochel and Johnson, 1984). Excellent descriptions of the morphometric characteristics of Holocene alluvial fans in the southern Canadian Rocky Mountains have been provided by Kostaschuk *et al.* (1986), Jackson (1987) and Jackson *et al.* (1987) but detailed sedimentologic descriptions are lacking.

The exposed fan successions occur in two distinct geologic belts of the Rocky Mountains: the Front Ranges, composed mainly of limestone, dolostone and shale, and the Main Ranges, dominated by sandstone, conglomerate, shale and breccia. Paleofan sediments are described from several locations in each of the two ranges (Fig. 3-1). This allows for comparisons of fan sedimentation in areas with contrasting source rocks.

Exposed sections are up to 65 metres thick and one kilometre long (Fig. 3-2a) permitting analysis of sedimentary patterns along much of the length of the paleofans. In addition, the entire fan succession, between a basal unconformity over bedrock and overlying glacigenic deposits, is locally exposed allowing for study of changes through the depositional history of the fans. The most distal parts of the paleofan deposits are not exposed due to intense glacial and fluvial erosion in the valley bottom and, to a lesser extent, burial by younger fluvial and alluvial fan sediments.

The objectives of this study were to: 1) investigate the sedimentologic characteristics of the deposits in order to determine environments of deposition and 2) document lateral and vertical facies relationships reflecting, respectively, spatial and temporal changes in sedimentation patterns, 3) compare past depositional events with processes observed on modern alluvial fans in the area and in other analogous settings, 4) develop criteria useful for recognizing different flow types in sedimentary successions, and 5) investigate the influence of bedrock lithology and other controls on sedimentation within the fans.

Study Area

The Jasper region is characterized by mountainous terrain with high relief, commonly in excess of 1500 m. Mount Robson, at an elevation of 3955 m, is the highest peak in the Canadian Rockies and occurs just west of the study area. The Rocky Mountains consist of northwest-southeast trending thrust sheets of mainly Paleozoic and Precambrian carbonate and clastic strata, formed during Late Cretaceous-Early Tertiary orogenic events. The Front Ranges, east of the Pyramid Thrust, are composed of steep, southwesterly dipping, thrusted carbonate, shale and sandstone strata from upper Cambrian to Jurassic in age. The Main Ranges subprovince consists of thick, relatively flat lying, or gently dipping, thrust blocks with broad folds and late, extensional, normal faults. They are mainly composed of quartzose sandstones, shales, siltstones and carbonates of Cambrian and Precambrian age (Price and Mountjoy, 1970; Mountjoy and

Price, 1985). Due to the prevailing westerly winds, the Main Ranges generally receive more snow and rainfall than the Front Ranges but local climate is highly variable, depending upon the local aspect and topographic situation. Approximately 400-500 millimetres of precipitation falls annually at Jasper town site, about 30% of which is snow.

Paleo-alluvial fan deposits were mapped at several localities including the Amber Mountain, Tekarra Creek and Portal Creek sites in the Main Ranges and the Roche Bonhomme, Jacques Creek, Roche Miette, Pocahontas and Punchbowl Falls sites in the Front Ranges (Fig. 3-1).The best exposures occur at Amber Mountain and Roche Miette (Fig. 3-2) and detailed descriptions of the paleofan deposits were obtained mainly at these two localities.

The drainage basin above the Roche Miette study site (Fig. 3-3a) is underlain almost entirely by limestones and shales (Mountjoy, 1959). The steep upper slopes in the basin, such as those on the face of Roche Miette (Fig. 3-2b), consist of massive limestone of the Devonian Palliser Formation. The lower parts of the basin are underlain by rocks of the Fairholme Group, mainly argillaceous limestones of the Mount Hawk Formation and shale of the Perdrix Formation. The drainage basin above the Amber Mountain section (Fig. 3-3b) is entirely underlain by Hadrynian rocks of the Miette Group which includes sandstone, shale, siltstone, conglomerate and limestone (Price *et al.*, 1977). The lower part of the basin is mainly underlain by the Old Fort Point Formation (Lower Miette Group) consisting of slate and siltstone with variable amounts of limestone including a conspicuous intraformational limestone breccia (Mountjoy and Price, 1985). The upper part of the basin and the valley bottom are mainly underlain by Middle Miette Group grits and sandstones with interbedded shales and siltstones.

Description and Classification of Coarse Clastic Deposits

Coarse clastic sediments, particularly unsorted or poorly sorted alluvial and glacigenic deposits, are difficult to describe and classify. Statistically accurate descriptions of these coarse clastics can rarely be obtained because detailed granulometric analysis are not practical (due in part to the large sediment volume required), visual field estimates of grain size distribution are not reliable and the textural characteristics of these deposits commonly are highly variable resulting in poor reproducibility. Secondly, many existing classification schemes of poorly sorted, coarse deposits have a genetic rather than descriptive basis (continually subject to reinterpretation) and therefore are of limited use in describing deposits in the field. This is particularly true of genetic classifications of sediment gravity flow deposits but also applies to classifications of glacial deposits (*e.g.* Boulton, 1976; Dreimanis, 1989; Goldthwait and Matsch, 1988) as discussed by Levson and Rutter (1989a). Classifications of flowing sediment-water mixtures have evolved rapidly in recent years and include schemes that are based mainly on flow rheology and particle support mechanisms (*e.g.* Middleton and Hampton, 1976; Lowe, 1979, 1982; Nardin et. al., 1979; Pierson and Costa, 1987). These classifications provide an important basis for interpreting processes but, since it is mainly the depositional characteristics that are preserved in the geologic record, field classification systems need to be based on sedimentologic criteria rather than inferred sediment transport and support mechanisms. For example, Postma (1986) emphasized that the deposits of several types of sediment gravity flows, including liquefied, fluidized, grain and modified-grain flows (Lowe, 1982), all have similar depositional characteristics and therefore can be classified as one group (cohesionless debris flows). Similarly, Smith (1986) noted that hyperconcentrated flood flow deposits should be identified by depositional characteristics rather than inferred support mechanism (Lowe, 1979) or empirical data on sediment concentration during transport (Beverage and Culbertson, 1964; Weirich, 1989).

Since the grain-size distribution of many poorly sorted coarse clastic deposits is both difficult to determine and highly variable, even within single beds, a term that encompasses a broad textural spectrum is needed. The term diamicton was introduced by Flint et al. (1960) to describe nonsorted or poorly sorted sediment that contains a wide range of particle sizes and this definition was endorsed by the International Quaternary Association Commission on the genesis and lithology of Quaternary deposits (Dreimanis, 1982; Dreimanis and Lundqvist, 1984). 'Diamictite' describes lithified equivalents of diamicton and 'diamict' is a general term applying to both unconsolidated and consolidated varieties (Frakes, 1978). Diamicton is a descriptive (non-genetic) term and its application does not require detailed and often impractical textural analyses, as is the case with more specific terms used in other classification schemes (*e.g.* Folk, 1980). It is thus a convenient field term and its proliferation in the recent literature is evidence of this utility.

Diamicton versus Gravel Lithofacies

Although diamicton has been long employed by Quaternary geologists to describe unsorted or very poorly sorted glacigenic sediments, Miall (1983) suggested that the term could also be applied to poorly sorted fluvial and alluvial deposits. He noted that this usage might generate some confusion in the distinction between gravels and diamicton, the two types of deposits exhibiting substantial overlap, and suggested that criteria for field discrimination of the two could be based on sorting characteristics. The two types of deposits are distinguished here by the presence or absence of mud in the matrix, with diamicton used only to describe unsorted or poorly sorted deposits containing silt and/or clay, with varying proportions of sand and gravel. Thus sandy gravel, with no mud in the matrix, is not referred to here as diamicton, even if matrix-supported. This is a useful field distinction, as silt and, particularly, clay in the matrix can readily be detected by hand-texturing and therefore can be used, with other characteristics, to distinguish different diamicton subfacies. It also has interpretive significance since clay content directly effects the rheological properties of debris flows (Hampton, 1975; Rodine and Johnson, 1976; Nemec and Steel, 1984; Costa, 1988; Weirich, 1989). This usage of the term diamicton is more restrictive than, but consistent with, that of Flint et al. (1960) and Dreimanis (1982) and is independent of both framework type (clast or matrix support) and actual proportions of different grain-size classes as recommended by Frakes (1978). Further restrictions on the definition of diamicton, for example by arbitrarily defining a minimum proportion of silt and clay such as 10% (Shultz, 1984; Eyles, 1987), are not used here as accurate determinations of silt and clay content, requiring laboratory analysis of each bed, are impractical especially in partially lithified and lithified deposits (Nemec and Steel, 1984).

MODERN ALLUVIAL FANS

Since depositional processes on modern alluvial fans in the study area are of relevance to genetic investigations of the paleofan sediments discussed below, a brief description of these processes is provided here. Observations were made on a number of modern fans in the region, particularly those in the vicinity of the detailed study sites at Amber Mountain and Roche Miette (Figs. 3-2 and 3-3). Incision of the paleoalluvial fan sediments at these sites by modern streams has resulted in the development of younger fans at lower elevations.

The shape of drainage basins feeding modern alluvial fans in the study area is largely controlled by the bedrock geology. Basins in the Front Ranges are confined between steeply dipping, northwest striking, thrust blocks and tend to be narrower than those in the Main Ranges. The drainage areas above the main feeder channels of modern alluvial fans at Amber Mountain and Roche Miette (Fig. 3-3) have lengths of about 3 kilometres and 4.25 kilometres, respectively (Table 3-1). The basin on Amber Mountain is approximately twice as wide as that at Roche Miette. Both drainage areas are otherwise similar in relief and planimetric area. Since the drainage basin characteristics in the Main and Front Range subprovinces are bedrock controlled, basins feeding the paleoalluvial fan successions at these two sites were probably similar to the modern basins (Fig. 3-3).

Sedimentation in fanhead channels above the intersection point on many modern alluvial fans is restricted mainly to isolated levee deposits and to channel gravels that are only temporarily stored in the channel with erosion and transportation occurring in subsequent events (Fannin and Rollerson, 1993). Sedimentation occurs largely below the intersection point where gradients decrease or channels widen. Areas of recent sedimentation are indicated by the absence of vegetation, with inactive parts supporting trees. Sedimentation is focused almost entirely along intermittently active channels, usually along one main system. These main channels are generally incised several metres in their upper reaches. They shallow and widen down fan but seldom exceed about 50 metres in width. At the modern Amber Mountain fan the main channel becomes less well defined down fan but at Roche Miette the main channel is more consistent in width and distinctness throughout the fan length.

A schematic cross-section through an area of recent sedimentation along the upper part of one of these channels on the Amber Mountain fan is shown in Figure 3-4. Several environments of active sedimentation are recognized. The channel directly below the fan apex has incised several metres through older fan deposits, now exposed along the north wall (Fig. 3-5a), and sedimentation is confined to coarse gravels on the channel floor. The bulk of recent sedimentation has occurred just down-fan of the incised channel on a low, crudely channelized, gravel flat, locally floored by gravelly diamicton (Fig. 3-5b). The latter is clast or matrix supported, matrix filled and dominated by large pebbles and cobbles. The massive structure, poor sorting and presence of boulder levees on the lateral flanks of the diamicton (Fig. 3-4) suggests deposition by debris flows (*c.f.* Lowe, 1979; Blair, 1987; Costa, 1988). Depositional events are largely non-erosive (Fig. 3-5b). Small, indistinct, abandoned channels incised into the debris flow deposits (Fig. 3-4) contain moderately to well sorted, imbricated, pebble gravels and minor patches of usually open work, cobble-lag deposits. The spatial association of these channels with the debris flow deposits suggests that they were cut by fluvial action in the late stages of the flow. A zone of more frequent fluvial activity lies approximately 50 centimetres below the adjacent gravel terrace (Fig. 3-5b). Small, seasonally active, braid channels in this zone are floored by mainly open work, imbricated, pebble and cobble gravels (Fig. 3-5c).

The area of recent sedimentation on the modern Amber Mountain fan is flanked by boulder levees up to several metres wide and a few metres high (Fig. 3-5d). Just below the fan apex the levees consist of boulders sitting on bedrock (Fig. 3-5e). Finer materials probably were removed by subsequent erosion or, alternatively, deposition in the proximal parts of the levees may have been restricted mainly to coarse boulders. Exposed sections of the levee deposits elsewhere reveal a layer of open work gravels, several clasts thick, overlying poorly sorted, matrix-filled gravelly diamicton (Fig. 3-5f). The presence of fines in the levee deposits at depth only, suggests that winnowing of the surface gravels occurred as a result of washing by runoff or rainwater following debris flow deposition.

The main levee, on the unconfined south side of the channel directly below the fan intersection point, is locally breached by small lobate deposits of cobble to boulder gravels that extend into the forested area beyond the levee (Fig. 3-4). The proximal sides of the lobes are characterized by cobble to boulder gravels with a sorted sandy pebble matrix (Fig. 3-5g) whereas the distal sides are open work (Fig. 3-5h). These lobate features are interpreted to be noncohesive debris flow or grain flow deposits formed by flows that overtopped or breached the levee. Similar deposits, showing vertical and horizontal sorting with surficial and distal parts of flow lobes characterized by coarser clasts, were described from a small alpine fan by Mears (1979). This interpretation is supported by the lobate form of the deposits, lack of fines in the gravels, crude imbrication and clast-support. A winnowed, cohesive debris flow origin is also possible but less likely as deposits of winnowed fines were not observed downstream of the lobes and the distribution of sorted matrix gravels indicates that fluvial reworking and deposition was restricted mainly to the proximal sides of the lobes. Similar modern debris flows with low clay contents have been observed in the southern Rocky Mountains of British Columbia (Jackson *et al.*, 1989).

Lobate deposits dominated by cobble to boulder gravels are common elsewhere in the Rocky Mountains on modern alluvial fans with clay poor source areas. One such fan, with a limestone dominated basin, along the North Saskatchewan River in northern Banff National Park was observed shortly after a mass movement event that temporarily blocked Highway 93. The flow formed a broad lobate deposit with its surface consisting of clast supported, cobble to boulder gravels with little interstitial matrix (Fig. 3-6a). The surface of the deposit was flat suggesting a low shear strength in contrast to more viscous debris flows. The highway was buried to a relatively uniform depth of about 2 metres across the width of the flow. The surficial openwork gravels were typically only a few clasts thick and matrix content generally increased with depth suggesting that deposition may have occurred as a cohesive debris flow with fines subsequently winnowed from the surface deposits by runoff and rainfall. However, some openwork gravel deposits up to 1 metre thick occurred locally and may have been noncohesive debris flow, grain flow or even sieve deposits.

Silt and clay deposits are rare on the modern fans, being restricted mainly to small low areas such as abandoned channel remnants, presently cutoff from stream flow. Muds settling from ponded water in these areas are thin, drape over underlying deposits and typically have desiccation cracks (Fig. 3-6b). Although the preservation potential of these deposits is likely to be low, gradual reactivation of abandoned channels can locally allow for deposition of fluvial gravels over fine grained sediments with little associated erosion (Fig. 3-6c).

Exposures of the modern fan sediments are generally shallow but typically reveal massive or crudely bedded, unsorted to poorly sorted, muddy or gravelly diamicton (Figs. 3-5a and 3-6d). Sand and gravel units occur mainly as thin trough-shaped lenses. Loess deposits interbedded with diamicton and gravel are rare in the fan successions except in the upper one or two metres. Near Disaster Point a number of silt and fine sand beds interpreted as loess occur, along with several buried Bm and Ah soil horizons and numerous charcoal fragments (Levson and Rutter, in press, their Fig. 3-11).

From the above observations the following general conclusions on recent fan sedimentation in the study area can be made: 1) episodic debris flow events deposit mainly poorly sorted, massive, gravelly diamicton; 2) debris flow levees consisting of open work, bouldery gravels underlain by gravelly diamicton are common but are laterally restricted to zones several metres wide; 3) debris flow deposits are commonly eroded and winnowed by intermittent stream flows resulting in deposition of weakly channelized, braided stream gravels and local cobble-lag concentrations; 4) silt and clay deposits are rare and usually aerially restricted; 5) noncohesive debris flow, grain flow and even sieve deposits are difficult to distinguish from winnowed cohesive debris flow deposits without subsurface exposures; 6) loess deposits and buried soils are restricted mainly to a thin surficial cap on the fan successions and are rarely interbedded with the gravelly deposits; 7) lobate noncohesive debris flow or grain flow deposits occur just below the intersection point and may breach older levees; 8) debris flow deposits are common in inactive parts of the upper fan and in medial fan reaches below the intersection point; 9) fluvial channel deposits may occur throughout the fan.

STRATIGRAPHIC SETTING OF THE PALEOFAN SUCCESSIONS

The location of measured sections in the study area is shown on Figure 3-1 and composite stratigraphic columns of measured sections from each of the main sites are given in Figure 3-7. Five main stratigraphic units are recognized (Levson and Rutter, in press). The oldest deposits (Unit 1, Fig. 3-7) consist mainly of locally derived, coarse-grained gravel and gravelly diamicton interpreted as paleofan deposits. These sediments are the main focus of this chapter. They are overlain by a proglacial succession dominated by glaciofluvial gravels and sands (Unit 2, Fig. 3-7) and glaciolacustrine silts and clays (Unit 3) at some sites. Till and associated glacigenic debris flows deposited during the last glaciation are represented by a thick sequence of matrix-supported diamicton beds (Unit 4). The upper-most unit (Unit 5, Fig. 3-7) consists mainly of glaciolacustrine and glaciofluvial sediments deposited during deglaciation, paraglacial debris flow deposits and minor loess. The paleofan sediments can be lithologically distinguished from the overlying glacigenic succession by the relatively high percentage of distally-derived clasts in the latter.

Chronology

The age of the paleoalluvial fan sediments is constrained by the radiocarbon chronology of stratigraphically younger glacigenic units in the region (Units 2 and 3). Wood dated at 29,100±560 (GSC 3792) in glaciofluvial gravels below till in the Jasper town site area provides a maximum age for the onset of the last glaciation (Levson and Rutter, 1989b). Glacial sediments that stratigraphically overlie the fan gravels are correlative with this till and therefore were deposited during the Late Wisconsinan stadial. Since deposits in the upper part of the fan successions are locally interbedded with glaciofluvial gravels (Figs. 3-7 and 3-8), alluvial fan sedimentation must have continued until the onset of glaciation in the Late Wisconsinan. Dateable materials within the fan sediments are rare but wood (Picea sp.), obtained from the Amber Mountain Section (Fig. 3-9), was dated at >39,750 (AECV 450c) and >48,000 (GSC-4646HP). Since the wood occurs relatively high up in the section, a large part of the fan succession at that site must have accumulated prior to 48 ka. In addition, as ice did not reach the area until sometime after 29 ka, fan sedimentation must have continued for at least an additional 20, 000 years after deposition of the wood bearing horizon. Thus, sedimentation in the Amber Mountain fan probably spanned all of the Mid-Wisconsinan interstadial and fan construction likely began in, or prior to, the Early Wisconsinan. The age of the oldest fan deposits is unknown but the presence of glacial erratics throughout the successions indicates that they began to accumulate sometime after Pleistocene glaciers first reached the area. The poor lithification and cementation of the deposits is also consistent with a Pleistocene age. Evidence suggestive of greater antiquity, such as tectonic disruption or Tertiary fossils, has not been found.

LITHOFACIES DESCRIPTIONS AND INTERPRETATIONS

Lithofacies Designations

The alluvial fan deposits described in this chapter include a wide variety of poorly sorted, massive gravels and diamictons deposited mainly by sediment gravity flows and other related flow types. The lithofacies classification presented here (Table 3-2) was developed to account for the relative abundance of these gravel-dominated deposits. The distinction of lithofacies in fluvial sand and gravel deposits is well established in the literature and generally follows that first proposed by Miall (1977, 1978). Miall's classification is based on dominant grain-size and sedimentary structures, principally stratification type. Additions to this scheme are required in order to accommodate the variability found in other sedimentary environments, particularly where poorly sorted, nonstratified deposits dominate such as in glacigenic, volcanic and alluvial fan successions. Thus, the classification presented here follows Miall (1977, 1978) but includes modifications such as those introduced by Smith (1987) for interstratified volcanic and sedimentary rocks, Walton and Palmer (1988) for lahar deposits, Eyles *et al.* (1983) and Miall (1983) for glacial deposits, and Waresback and Turbeville (1990) for volcanogenic alluvial fan successions.

The sedimentary characteristics of gravelly deposits in the alluvial fan successions result from a wide spectrum in flow rheology varying from normal stream flow to cohesive debris flow. Since deposits of intermediate flow types between these two end-members may show similar sedimentary characteristics, they often can not be readily distinguished. Hyperconcentrated flood flow deposits, for example, are distinguished from cohesive debris flow deposits by clast support, fabric type and normal grading (Smith, 1986) but their distinction from noncohesive debris flow deposits is less clear. Similarly, the distinction between the deposits of hyperconcentrated flows and high velocity stream flows with high sediment concentrations is not always obvious (Morison and Hein, 1987). For these reasons, the genesis of many poorly sorted gravelly beds can not be precisely determined, especially in the absence of detailed clast fabric data and complete grain-size analysis of both matrix and clasts. Since bed-by-bed analysis using these methods is often impractical, it is necessary to rely on other sedimentologic characteristics of the deposits for lithofacies designations.

The characteristics used here to distinguish different poorly sorted gravel and gravelly diamicton deposits in the field include: grading (normal, inverse or ungraded and coarse-tail versus distribution grading), stratification (planar and trough cross-bedding, horizontal bedding), imbrication, clast framework (clast-supported versus matrix-supported and openwork versus matrix-filled), bed thickness, sorting, matrix-texture, clast-size (mean and maximum) and clast roundness. Detailed data on grain-size distribution, fabric type (unimodal, bimodal and polymodal) and imbrication type (a[p], a[i] and a[t], b[i] or both) (terminology of Harms *et al.*, 1982), where collected, are used mainly to make genetic and rheologic interpretations of the different lithofacies. In all cases, only descriptive (non-genetic) criteria, readily observable in the field are used in the lithofacies designations. However, some sedimentary characteristics, such as stratification, clast imbrication and grading, can also be used to infer flow type and, where present, they are preferred over characteristics that do not reflect rheologic or depositional processes. In the absence of such characteristics, as is common in many massive beds of poorly sorted gravel and diamicton, deposit genesis can not always be precisely determined.

The most critical factors for determining flow rheology are clast-orientation, grain-size data (particularly maximum clast-size and clay/silt content of the matrix), grading, stratification and framework type. The proportion of matrix to clasts and mud content are especially important factors controlling mode of sediment transport and deposition in flows intermediate between stream and debris flows (Smith, 1986). The rheological properties of the flows, however, are also dependent on both the composition and quantity of clay and on the absolute sizes of the clasts (Rodine and Johnson, 1976). Thus characteristics such as sorting and mean or maximum clast-size can not be used independently to make inferences regarding flow type. Nevertheless, as well illustrated by Nemec and Steel (1984), basic sedimentologic data such as bed thickness and maximum clast-size, in conjunction with other data, can be helpful in making genetic inferences. These data are included here in the following descriptions but they are not used as criteria for defining lithofacies.

Clast-Supported Diamicton (Dc)

Deposits of this facies volumetrically dominate the fan successions (Figs. 3-7 to 3-10). The facies is characterized by clast-supported, matrix-filled, unsorted or very poorly sorted, large pebble to boulder diamicton (Fig. 3-11a). Deposits of this facies are highly heterogeneous and only characteristics common to most beds are described here. Descriptions specific to three subfacies, recognized mainly on the basis of differences in grading type and sedimentary structure (Table 3-2), are provided separately below. Beds of this facies average about 0.5 - 1.5 metres in thickness with a range from 0.2 - 4.8 metres. Bed boundaries are gradational and poorly defined, commonly distinguished only by the relative abundance of cobbles and boulders or by other gradual changes in the grain-size distribution of the deposits (Figs. 3-11a, b). As a result, it is difficult to recognize the presence or absence of grading and to trace individual beds laterally. However, most beds appear to be ungraded and from at least a few metres to tens of metres wide. Beds are generally thicker and more continuous than, and may grade laterally into, massive or stratified gravel facies.

Matrix textures are polymodal, typically in the muddy-sand range (70-80% sand, 20-30% mud) with low proportions of clay (Fig. 3-12) especially in relation to the gravel fraction. Maximum clast size is generally in the large cobble to boulder range, the latter being common in the most poorly sorted beds. Angular and subangular clasts are abundant. Some very angular clasts are of the same rock type as bedrock at nearby exposures directly up-valley. Outsized clasts are common and beds may be matrix-rich or clast-rich. Spatial variations in clast content within beds are common with some clast-rich beds interbedded with, or grading laterally into, matrix-rich strata. In addition, some beds, up to 2 metres thick, consist of cobble or boulder clasts 'floating' in a disorganized silty sand-pebble matrix (Fig. 3-11c). The granular to small pebble fraction in coarse gravelly deposits of this type may function as interstitial matrix (Wells, 1984).

Clasts in this facies display either crude a-axis (ap ai) or, to a lesser extent, b-axis (at bi) imbrication. Fabrics are generally poorly developed and clast orientation plots in gravelly diamicton beds show a significantly higher degree of scatter than those in massive to crudely stratified gravel facies (S1=0.59 vs. 0.86; Fig. 3-13).

Discontinuous lenses of silty diamicton, sand or granule to pebble gravel are locally interbedded with deposits of this facies. Clast-rich beds are commonly interbedded with massive, pebble to cobble, gravel facies (Gm). Bed contacts are indistinct except contacts over sandy strata which are often sharp, loaded and unconformable.

Subfacies Dci, Dcn and Dcs

Although most beds of this facies are ungraded, gradual vertical changes in grain-size distribution, particularly in the abundance of coarse clasts, allows for the distinction of two subfacies. These include beds with inverse grading (subfacies Dci) and rare beds with coarse-tail normal grading (subfacies Dcn). The latter is indicated by concentrations of cobbles and boulders in the lower parts of the beds. Coarse-tail inverse grading (subfacies Dci) is locally present in strata with relatively high clast concentrations and in cobble or boulder-rich beds. This subfacies locally shows inverse grading at the base of beds (Fig. 3-9) that otherwise are ungraded or grade upwards into subfacies Dcn. These beds rarely are stratified at the base and they may also be successively overlain by gravels (Gm), granular to pebbly sands (Sm) and finally massive silts and fine sands (Fl, Fm) (see, for example bed at base of Amber Mountain mid-fan Section; Fig. 3-9). Horizontally-bedded gravels may also cap these successions.

A third subfacies (Dcs) is recognized by the local development of very crude stratification and weak imbrication. Subfacies Dcs typically has a very low proportion of clay in the matrix. Disorganized clusters of large clasts also occur locally in subfacies

Dcs, particularly along large-scale scoured surfaces (Figs. 3-8, 3-11a). Clast size within the clusters usually decreases up-fan with the largest clasts occurring on the down-flow end. In the lee of the clusters poorly sorted beds of finer gravels may also occur. Clasts in the clusters may show a weak imbrication (Fig. 3-11a).

Interpretation

Clast-Supported Diamicton (Facies Dc)

Variations in grain-size distribution and sedimentologic character between and within beds of this facies and associated subfacies indicate a broad spectrum in rheology and depositional processes. Poor sorting, large maximum clast-sizes and thick beds indicate high energy flows that rapidly deposit large volumes of sediment. As discussed further below, poorly sorted gravelly sediments of this type may be deposited by any of a variety of flow types intermediate between fully turbulent stream flows and cohesive debris flows including noncohesive debris flows or pseudoplastic debris flows and hyperconcentrated flood flows (Shultz, 1984; Postma, 1986; Smith 1986, 1987; Massari *et al.*, 1993). These flow types can be differentiated in some cases on the basis of sedimentologic characteristics, principally grading type and, to a lesser extent, stratification, imbrication and matrix clay content.

The presence of angular clasts, weak fabrics, poor sorting and a massive, ungraded structure in beds of facies Dc suggests that they are debris flow deposits. Massive, poorly sorted sediments with a disorganized fabric and gradational bed contacts are typical of subaerial debris flow deposits (Fisher, 1971; Bull, 1972; Harvey, 1984; Kochel and Johnson, 1984; Wells, 1984). Similar deposits with few fines and a massive, ungraded character were interpreted by Larsen and Steel (1978), Burgisser (1984) and Massari *et al.* (1993) as noncohesive gravelly debris flows. The low matrix content in these beds also suggests deposition from flows in which clasts remain mostly in contact with each other. In gravelly flows of this type particles are suspended by buoyancy and clast collisions rather than by the cohesive strength of the matrix (Pierson, 1981; and Burgisser, 1984). The general absence of grading may be explained by syndepositional mixing and churning and/or rapid deposition (Hein, 1982; Wells, 1984). Nemec and Steel (1984) reported that well developed normal or inverse grading are uncommon in subaerial debris flows.

The presence of small amounts of clay in the matrix of most beds may also account for the lack of grading in this facies and suggests deposition by flows that were transitional between cohesionless and cohesive debris flows. Rodine and Johnston (1976) provided theoretical and experimental evidence to demonstrate that, even in poorly sorted deposits in which clasts form up to 95% of the total volume, low concentrations of clay can allow for frictionless flow, with apparent cohesion determined mainly by the water content of the fluid phase. In flows of this type, where clasts are largely in contact and not totally suspended in the matrix, the resulting deposits show mainly clast-supported frameworks (Lowe, 1982). Similar, massive, clast-supported diamictons in alluvial fan successions in the Cutler Formation were interpreted by Shultz (1984) as pseudoplastic debris flow deposits, characterized by the absence or extreme reduction of a rigid plug and low matrix strength. The composition and amount of clay in clast-supported (and matrix-supported) diamictons is a critical factor in determining flow type as clay both enhances buoyant lift and acts as a lubricant, reducing frictional locking (Rodine and Johnston, 1976; Lowe, 1982).

Jackson *et al.* (1989) described deposits similar to this facies from a modern debris flow fan at Cathedral Mountain in the southern Canadian Rocky Mountain Main Ranges. Debris from an event in 1984, which flowed at several metres per second and produced an abundant spray of silty water, was found to consist of 30-40% cobbles and boulders, up to 1 m in diameter, in a gravel matrix with little sand and silt and only trace amounts of clay (Jackson *et al.*, 1989). Similarly, debris from an event in 1978 at the same site consisted of 60-70% gravel with clasts up to 5 m in diameter. Jackson (1987) also found that the clay content in the minus 64 mm fraction, from numerous debris flow fan deposits throughout the Rocky Mountains, was generally less than 5%. These data indicate that noncohesive debris flows are a widespread phenomenon on Rocky Mountain alluvial fans.

Inversely Graded Subfacies (Dci)

Coarse-tail inverse grading in this subfacies suggests that dispersive pressure due to grain-to-grain collisions was a dominant grain support mechanism. Shultz (1984) stressed the importance of dispersive pressure for the development of inverse grading in inertial flows (where grain collisions predominate over viscous granular interactions) even in poorly sorted sediment dispersions. Coarse-tail inverse grading probably results from larger clasts moving upwards in the flow to regions of less shear stress (Bagnold, 1954) or possibly as a result of kinetic sieving where smaller clasts move down between larger clasts (Middleton, 1970; Walker, 1975). The paucity of fines, high clast content and weak imbrication in these beds suggest continuous clast collisions, with dispersive pressure responsible for development of the inverse grading.

Inverse grading in thin basal zones of this subfacies may develop when the concentration of sediment at the bottom of a flow increases due to clast settling (*c.f.* Lowe, 1976, 1982). These basal particle layers, referred to as traction carpets, are maintained by dispersive pressure which suppresses turbulence and causes the development of inverse grading. With continued fallout of sediment from the overlying flow, the traction carpet layer may collapse and freeze resulting in preservation of an inversely graded basal zone (Lowe, 1982). The rarity of inversely graded basal beds in this subfacies may be due to the dominance of gradational bed contacts making recognition of these thin zones difficult. Their use in interpretation is thus usually restricted to beds with sharp contacts that, for example, overlie fluvial sediments interstratified with the debris flow deposits. Smith (1986) also noted that inversely graded basal beds are rare in clay poor gravelly deposits.

Although normally restricted to the basal shear zone in cohesionless flows, thicker inversely graded layers may develop due either to successively upward freezing of basal inertia-flow layers or to changes in slope or roughness producing shear at higher levels in the flow (Lowe, 1982; Walton and Palmer, 1988). Lowe (1982) noted that in density-modified grain flows, typically consisting of clast-supported gravels with a poorly sorted matrix, inversely graded units are generally thicker than 40 centimetres. Inverse grading in this subfacies is best developed in cobble or boulder dominated beds and poorly developed or absent in pebbly beds, reflecting the direct proportional relationship between grain size and dispersive pressure (Bagnold, 1954).

Beds of this subfacies that fine upwards into normally graded diamicton beds (subfacies Dcn) as well as gravel or gravelly sand beds, may reflect the influence of turbulence in the later stages of flow. Capping stratified granule to pebble gravel beds and fine sand and silt beds are interpreted, respectively, as traction and suspension sediments deposited during the more fluid waning stages of the flow or during intermittent fluvial activity between debris flow events. Lowe (1979, 1982) inferred that similar inversely to normally graded deposits formed from suspension deposition subsequent to frictional freezing of underlying traction carpets. These deposits are equivalent to the inversely to normally-graded, conglomerate facies of Walker (1975, 1977), Aalto (1976) and Massari (1984). They are usually characterized by a(p) a(i) fabrics. Clasts with a preferred orientation of long-axes parallel to flow, often with an upstream dip have been documented at numerous sites in inverse to normally graded conglomerates (Walker, 1975; Massari, 1984). They probably were deposited by cohesionless debris flows on steep slopes where applied basal shear stresses are high enough to maintain dispersive pressures.

Inversely to normally graded, clast-supported deposits similar to some beds in this subfacies were also described in lahars at Mount St. Helens (Scott, 1986) and in the Mount Dutton Formation (Walton and Palmer, 1988). They were interpreted as basal flow, 'whaleback' bar deposits, that form at points of lateral spreading of flow and at other sites of decreased flow competence. Although uncommon in lahar channel facies (Scott, 1986), these deposits may be more common on alluvial fans where lateral spreading of flow frequently occurs at the down-flow end of fan channels (*c.f.* Fannin and Rollerson, 1993).

Normally Graded Subfacies (Dcn)

The influence of turbulence, reflected by normal grading is rarely seen. Beds of this subfacies generally contain less fines than ungraded or inversely graded beds and they may have been deposited by hyperconcentrated flood flows. The deposits of hyperconcentrated flood flows may grade laterally or vertically into cohesionless debris flow deposits and they may not always be distinguishable, particularly in the absence of detailed fabric data. The presence of bimodal fabrics in some ungraded beds suggests that, in addition to relatively rare normally graded beds, some massive beds may also have been deposited by hyperconcentrated flood flows (see facies Gm(a) below). Normal grading in beds of this subfacies may also have developed as a result of clast settling in pseudoplastic debris flows having relatively high water contents (Shultz, 1984). However, these low yield strength, 'slurry' flows may be distinguished from hyperconcentrated flood flows by a higher content of fines.

Stratified Subfacies (Dcs)

The local development of very crude stratification, weak imbrication and the dominance of clast-supported textures in this subfacies suggest a fluidal (cohesionless) rather than plastic (cohesive) flow behavior during the depositional phase (Postma, 1986). The presence of clast-clusters and their local association with scoured surfaces indicate that turbulence and tractional deposition occurred locally in the development of some beds, probably during more fluid stages of flow at peak discharges. Shultz (1984) suggested that clast-supported diamictons, similar to this subfacies, occupying channel-like depressions in the Cutler Formation were formed by debris flows that were turbulent during early, high-velocity, stages of the flow. Deposition of clast clusters in this subfacies appears to have occurred following high discharge events. Similar, disorganized to weakly imbricated clast clusters, were also described in poorly sorted, gravel bed, alluvial channels and flume experiments by Brayshaw (1984). The cluster bedforms developed during falling stage from flood discharges, subsequent to the deposition of large clasts (median size > D95, *i.e.* coarser than 95% of the bedload) that formed flow obstacles. Fine grained (D8 to D46) wake accumulations were deposited in the obstacle shadow and comparatively coarse (D74 to D94) clasts accumulated in a cluster on the stoss-side (Brayshaw, 1984).

'Outsized' Clasts

Large cobble and boulder clasts, common in facies Dc and subfacies Dci, Dcn and Dcs, may occur as a result of transport along rheologic interfaces between basal inertia-flow layers (traction carpets) and overlying, more rapid, turbulent layers, known to occur in experimental sediment gravity flows (Postma *et al.*, 1988). These outsized clasts, probably initially suspended by high turbulence in early phases of flow, glide along these interfaces and develop a(p) a(i) fabrics when they are transported mainly within the top of inertia-flow layers where laminar shear is high (Rees, 1968, 1983; Allen, 1984). Clasts with a(t) b(i) orientations probably were dragged on top of the underlying inertia-flow layer, with their centre of gravity above the interface in the overlying turbulent flow layer. Postma *et al.* (1988) suggested that distinct rheologic interfaces, along which outsized clasts may be transported, can develop in both subaerial and subaqueous sediment gravity flows as well as within subaerial hyperconcentrated flows. In the latter, a rheologic interface may develop as suspended particles become increasingly concentrated towards the bed during flow deceleration resulting in a higher viscosity suspension that will suppress turbulence and become pseudolaminar. Large clasts may retain their a(t) b(i) fabric in hyperconcentrated flows when laminar shear in the basal zone is only sufficient to reorient smaller clasts (Postma *et al.*, 1988).

Matrix-Supported Diamicton (Dm)

Matrix-supported diamicton beds, uncommon within the fan successions, are generally structureless (massive) and very poorly sorted to unsorted. Typically, a high proportion (40-50%) of the matrix consists of silt and clay with measured clay contents in the matrix as high as 30 per cent (Fig. 3-12). Clasts tend to have preferred a-axes orientations parallel to inferred paleoflow direction (bedding dip-direction). The shape of clasts in the diamictons varies widely from angular to rounded. Large clasts, up to very large boulder-size, are common. Gravel-sized clasts (>2 mm) generally compose much less than 50% of the total volume of the deposits. Matrix-rich beds (<15% gravel) locally are gradationally overlain by silts and commonly occur as lens-shaped intrabeds, 10 to 100 centimetres thick and 2 to 10 metres wide, although more extensive beds of this type are also present.

Bed contacts are usually gradational but over sandy strata they may be unconformable and show load structures. Matrix-supported diamicton beds are commonly interbedded with stratified gravel and sand facies.

Subfacies Dmn and Dmi

Most beds of facies Dm are ungraded but two subfacies are indicated by the rare presence of coarse-tail normal grading (subfacies Dmn) or inverse grading (subfacies Dmi). Normal grading is indicated mostly by basal concentrations of cobbles and boulders. Where inverse grading is present, it is generally weak and occurs in the most clast-rich beds (> 35% gravel). Thin inversely graded basal layers, sometimes with diffuse laminae, indicated by slight textural variations in the diamicton, also rarely occur.

Interpretation

Matrix-Supported Diamicton (Facies Dm)

Diamictons of this facies are interpreted as cohesive debris flow deposits. Modern debris flow sediments have similar characteristics (Bull, 1972; Kochel and Johnson, 1984) and structureless, poorly sorted, matrix-supported diamictons with non-erosional lower contacts have been interpreted as debris flow deposits by numerous authors (*e.g.* Harvey, 1984; Wells, 1984; Kostaschuk *et al.*, 1986; Smith, 1986; Walton and Palmer, 1988; Scott *et al.*, 1992; Whipple and Dunne, 1992; Nemec and Postma, 1993). Matrix support and relatively high clay contents suggest that the flows were viscous with a cohesive matrix. The dominance of a-axis parallel fabrics and non-erosive bed bases and the absence of traction induced stratification suggest laminar flow with strength of the matrix as the main sediment support mechanism. Cohesive debris flows of this type with high strength may develop a rigid plug at the top of the flow in a zone where the yield strength is not exceeded (Postma, 1986; Pierson and Costa, 1987).

Debris flows can transport material as both traction and suspended loads with large clasts, whose weight exceeds the combined effects of yield strength and lift forces, transported along the bed and smaller grains supported by various combinations of cohesive matrix support, dispersive pressure and flow turbulence (Lowe, 1979). The uniform dispersal of clasts throughout matrix-supported debris flow deposits, may reflect weak dispersive pressure or flow turbulence at some point prior to deposition (Lowe, 1979, 1982; Walton and Palmer, 1988).

Exceptionally large or outsized clasts, common in most beds of this facies, are a well documented characteristic of many debris flow deposits (Fisher, 1971; Hein, 1982; Lowe, 1982; Massari, 1984; Clifton, 1984; Kostaschuk *et al.*, 1986; Postma *et al.*, 1988; Nemec and Postma, 1993). Maximum clast size in cohesive debris flows is a reflection of flow strength (Hampton, 1975). Cohesive flows can transport large boulders, whose diameter may exceed flow thickness. Support of such large clasts is explained by rafting in a rigid plug with high strength (Postma, 1986). Some poorly sorted debris flows with low strength, allowing flow on low slopes, can still support large clasts as the high density of the debris may nearly equal the average density of individual boulders (Rodine and Johnson, 1976).

The lenticular cross-section and limited aerial extent of many matrix-rich diamicton beds of this facies, indicates that these flows were more viscous and confined to topographically low areas such as pre-existing channels. This interpretation is supported by the association of these beds mainly with channel gravel and sand facies.

Normally Graded Subfacies (Dmn)

Coarse-tail grading in a few beds of this facies may have resulted from clast settling during flow (Denlinger et. al., 1984; Shultz, 1984; Smith, 1986). Large clasts, whose weight exceeds the yield strength of the flow, may settle out and either be deposited or dragged along the bed (Postma, 1986). Lowe (1979) described clasts more than 0.5 metre in diameter that were rolled or, less commonly, shoved and dragged by modern subaerial debris flows. The transported clasts moved more slowly than the rest of the flows and protruded above the general surface of the flow. Normal grading in some beds may also reflect turbulence in an early phase of the flow. Grading in such flows, formed by settling of the coarsest clasts prior to flow transformation, may have been subsequently preserved by cohesive freezing triggered, for example, by the incorporation of eroded sediment (Enos, 1977; Lowe, 1982; Nemec and Steel, 1984). Normal grading in matrix-supported diamictons, interpreted as debris flow deposits with low to moderate strength by Walton and Palmer (1988), was attributed to a turbulent phase in the history of the flows.

Inversely Graded Subfacies (Dmi)

Inverse grading occurs in clast-rich beds of this facies and suggests that dispersive pressures may locally be important. Inverse grading may also form in clay-rich debris flows due to an upward-increasing yield strength allowing larger clasts to be supported in the upper part of the flow (Naylor, 1980). Since cohesive debris flows may develop from fully sheared flows or may be characterized by a rigid plug throughout their flow history with shear occurring only at the base of the flow, differences in grading type are commonly inferred to be inherited features reflecting variations in flow history (Lowe, 1982; Postma, 1986). Basal shear zones with inverse grading, for example, may develop in fully sheared flows when the shear rate drops below the yield strength at the top of the flow where the velocity gradient is lowest (Postma, 1986). This results in the formation of a rigid plug overlying a basal shear zone where thin, inversely graded, layers with diffuse shear laminae, such as are rarely present in this subfacies, may occur.

Massive to Crudely Bedded Gravels (Gm)

This facies consists of massive to crudely stratified gravels with horizontal or subhorizontal bedding. The gravels are typically clast-supported, moderately to well rounded and imbricated on their b-axes (at bi; Fig. 3-13). Beds are generally 5-50 centimetres thick and may be laterally traceable for up to 10 metres. Large pebble to cobble gravel beds are generally matrix-filled, more than 20 centimetres thick and poorly to moderately sorted. Boulders are rare. Poorly to well sorted, granule to medium pebble beds are less common and usually less than 20 centimetres thick. Clasts are generally subrounded to rounded. Those up to medium pebble-size are commonly tabular while larger clasts tend to be more equant. Sandy, matrix-filled gravels with few fines generally dominate over open work beds. Interbedded sand lenses are locally common.

Where present, stratification in this facies is poor, with gradual changes in grain-size distribution and sorting between beds (Fig. 3-11b) but is locally more pronounced where fine gravel or open work beds occur (Fig. 3-14a). Strata are commonly slightly trough-shaped and dip down-valley at low angles (typically 1° to 5°). Steeper dipping beds (up to 10°) are uncommon and are usually interbedded with thin, irregular trough-shaped sand lenses commonly with deformed bedding. Some matrix-filled beds, dominated by large pebble to cobble-sized gravels, are strikingly bimodal (Fig. 3-14b); matrix sands in these beds may have horizontal to wavy stratification, cross-laminae or contorted laminae. Sandy gravel beds commonly grade upwards into open work gravels (Fig. 3-14b) and some pebble gravel beds fine upwards.

Open work beds are typically stained by manganese and iron oxides and strongly enhance stratification where present (Fig. 3-14c, d). They occur mainly as discontinuous interbeds or thin trough-shaped lenses 1-5 metres wide, locally with abrupt lower contacts (Fig. 3-14d). These beds commonly consist of rounded, imbricated gravels, one or two clasts thick, that are generally coarser than those in adjacent gravelly strata and finer than in adjacent diamicton beds (Figs. 3-9 and 3-10). When closely spaced, they may appear as low-angle, trough cross-bedded sets (Fig. 3-14d). Laterally extensive beds ('sheets') of openwork gravels are rare. They are typically moderately well sorted and composed largely of angular fragments with a finer 'infiltration' matrix resulting in a bimodal grain-size distribution. One subfacies [Gm(a)] of facies Gm is distinguished on the basis of differences in fabric type, imbrication strength, matrix sorting, clast angularity, grading and clast-support.

Subfacies Gm(a)

Gravels of subfacies Gm(a) are characterized by poorly imbricated gravels with long axes of clasts oriented both transverse and parallel to flow (Fig. 3-15). They are generally poorly sorted and structureless but locally may be weakly stratified or normally graded. They typically are clast supported with a polymodal, sand-pebble matrix, although matrix-supported beds may also occur. Clasts are subangular to rounded and outsized, large cobble to small boulder clasts are common. Lower contacts may be gradational or erosional, as indicated by scouring. Subfacies Gm(a) is equivalent to that defined by Smith (1987) and his facies designation is retained here.

Rose diagrams produced from fabric analyses in beds of this subfacies display a bimodal distribution of long-axis orientations. The a-axes of many clasts are oriented parallel to inferred paleoflow direction but about 40 to 50 per cent have a transverse orientation (Fig. 3-15). The bimodal distribution of clast long-axes is controlled by clast size with boulders, cobbles and some very large pebbles exhibiting a-axis transverse-to-flow orientations and small to large pebbles displaying a-axis parallel-to-flow alignments.

Interpretation

Massive Gravels (Facies Gm)

Massive to crudely bedded gravels of this facies are interpreted as tractional channel gravels, deposited mainly in shallow bars and as channel lags or, rarely, in sieve lobes. Deposits of this facies are similar to facies Gm proposed by Miall (1978). Similar deposits on modern alluvial fans in the Front Ranges in the southern Canadian Rocky Mountains were also interpreted as stream channel deposits and longitudinal bar gravels by Kostaschuk *et al.* (1986). Imbrication and crude stratification in most beds indicate the dominance of tractional deposition. Crude horizontal to low-angle inclined stratification is characteristic of longitudinal bars and low relief diffuse gravel sheets (Hein and Walker, 1977; Hein, 1984). The low angle of bedding may be due to coarse bedload, shallow water depths, and/or high flow velocities at the time of deposition. These stream characteristics result in low relief bedforms without angle-of-repose foreset margins because of the low flow depth/grain size ratio or rapid deposition without subsequent reworking under lower flow conditions (Smith, 1985; Morison and Hein, 1987). Slightly steeper dipping beds may be related to deposition along the margins of diagonal bars (Harms *et al.*, 1982). Gravels in trough-shaped lenses with sharp lower contacts (Figs. 3-14d, e) are interpreted as channel lag deposits. The dominance of imbricated, well rounded, large clasts, open frameworks and scoured lower contacts are consistent with a channel lag origin. The dominant clast size in beds of this facies is generally smaller than in adjacent debris flow beds suggesting deposition during relatively low energy events.

The stratified sand matrix in some coarse gravel beds (*e.g.* Figure 3-14b) may reflect depositional processes in the lee of channel bedforms where the coarse gravel is supplied by bedload transport from upstream and the sands are simultaneously deposited from a separation eddy in the lee of the bedform. A detailed description of these processes was provided by Shaw and Gorrell (1991) to explain similar features found in gravel dunes formed by high velocity flows. Shared characteristics include bimodality of the deposits, sorting of the sandy matrix, large differences in the grain size of matrix and clasts, and laminae in the matrix sands. The upwards gradation into open work beds may be explained by an increase in gravel bedload supply relative to deposition from the backflow (Shaw and Gorrell, 1991) or by turbulence or other grain-support mechanisms keeping the sand-sized clasts in suspension (Walker, 1975). Alternatively, open work gravels at the tops of these beds may reflect winnowing of the matrix by subsequent stream activity unrelated to the depositional event. Deformation of matrix stratification in some gravelly strata may be due to loading or rapid deposition of gravels on sandy sediments, possibly as a result of pulsating flow conditions.

Where gravels of this facies overlie openwork gravels or other sediments with high permeability, they may have been deposited as sieve deposits. These deposits are clast supported, usually openwork, show crude stratification and imbrication, have gradational lower contacts, and commonly contain numerous angular clasts. Similar coarse gravels interpreted as sieve deposits have been observed on modern fans (Hooke, 1967; Bull, 1972; Wasson, 1974; Collinson, 1986; Kostaschuk *et al.*, 1986; Nemec and Postma, 1993) but they are rare in the paleofan successions described here. Sieve deposits form directly below the intersection point when water rapidly infiltrates into underlying permeable sediments resulting in the deposition of gravel lobes. They are common where source areas provide little fine sand, silt and clay to the fans (Bull, 1972). They occur both as extensive sheets and discontinuous lenses and they are distinguished from channel gravels by the relative abundance of angular clasts, poor sorting, weak imbrication and lack of cross-bedding (Harvey, 1984). Since sieve deposits may form from the early clear-water stages of floods (Wasson, 1974), they can have characteristics similar to other tractional deposits (such as channel lag and bar gravels) but they may be differentiated by the presence of gradational lower contacts and angular clasts. Sieve deposits can also be confused with winnowed debris flow deposits but they may be distinguished from them by features indicating traction transport such as crude stratification and imbrication. Blair and McPherson (1992, 1993), for example, reinterpreted massive fan gravels, described as sieve deposits by Hooke (1967, 1993), as winnowed debris flow deposits because they are matrix supported at depth and they have features characteristic of debris flows such as levees, lobes and a massive structure. Deposition by sieve mechanisms may occur only in parts of gravelly flows where suitable conditions exist such as where mud contents are low, water contents are high and/or where the substrate is highly permeable.

Subfacies Gm(a)

Deposits of this subfacies, particularly those dominated by cobble to boulder-sized gravels, are similar to hyperconcentrated flood flow deposits described by Smith (1986, 1987). Poor sorting and lack of stratification in the matrix and the overall clast-supported framework, suggest that suspension deposition of the matrix occurred simultaneously with traction sedimentation of the gravels. Pebbles and very coarse sands in the matrix are too coarse to have been deposited by post-depositional infiltration between the framework clasts and further indicate a component of suspension sedimentation. Hyperconcentrated flood flows are characterized by high sediment concentrations with the coarsest clasts transported by traction and smaller ones dispersed above the bed due to turbulence and grain interactions. Fines are kept in suspension by turbulence and for the most part are not deposited with the gravels but, due to increased viscosity, they may help support the larger clasts during transport (Smith, 1986). Turbulence, at least in the early stages of the flows, is also responsible for the scoured contacts that occur at the base of some beds of this subfacies.

Hyperconcentrated flood flow deposits are transitional between clast-supported, stratified and imbricated fluvial gravels and poorly sorted, matrix-supported, massive, debris flow deposits (Smith, 1986). They are distinguished from stream flow gravels by their poorly sorted polymodal matrix, poor imbrication, bimodal fabric, lack of associated sand lenses and great bed thickness. Postma *et al.* (1988) described similar deposits, referred to as bipartite turbulent suspensions, that form from fully turbulent sediment-water mixtures in which clasts become increasingly concentrated towards the base during flow deceleration. This may result in the normal grading that characterizes some beds of this subfacies.

The grain-size controlled, bimodal fabrics, characteristic of this subfacies (Fig. 3-15) may form as a result of settling of the coarsest fractions towards the base of the flow just prior to deposition. Reduced turbulence in this relatively high viscosity basal suspension may eventually allow for the development of pseudolaminar flow and laminar shear may selectively reorient the smaller clasts into long-axis parallel-to-flow positions but may be insufficient to realign larger clasts (Postma et al., 1988). If these inferences are correct, laminar shear may develop to different degrees under different flow conditions and, as a result, the minimum size of clasts retaining a b-axis transverse-to-flow fabric may vary from bed to bed. This may explain why some fabric results from this subfacies, unlike those described by Smith (1986), indicate that, in addition to cobbles and boulders, large pebbles also retain a b-axis imbrication.

Cross-Stratified Gravel Facies (Gt, Gp)

Gravels of this group are characterized by moderately to well developed cross-stratification. They include trough (Gt) and, less commonly, planar (Gp) cross-bedded gravels. These facies are rare in the alluvial fan successions, occurring mainly in Front Range fans, but they are common in glaciofluvial successions overlying the fan deposits (Fig. 3-7).

Clast support and well developed imbrication are ubiquitous characteristics of cross-stratified gravel facies. Clasts are imbricated on their b-axes with a-axes orientations generally transverse to flow direction [a(t) b(i) type]. Clasts are generally subrounded to well-rounded. Sandy, matrix-filled gravels generally dominate over open-work beds. Cross-stratified gravel facies commonly are gradationally interbedded with massive gravels or cross-stratified sands. Contacts between beds vary from sharp to gradational.

Trough cross-bedded gravels occur mainly in lens shaped units up to a few metres thick. They typically fine upwards into sandy facies and commonly contain diamicton intrabeds (*e.g.* Fig. 3-9). Planar cross-bedded gravels are rare in the fan deposits and usually occur within channelized gravel and sand units. Individual beds in planar cross-stratified gravels generally are less than one metre thick and commonly show normal grading.

Interpretation

Well developed stratification, good sorting, b-axis imbrication, and the presence of well-rounded clasts and open-work beds suggests that gravels of this facies were deposited by traction dominated fluvial processes. Rolling and saltation of clasts during bedload transport resulted in well-developed imbrication and well-rounded clasts.

The presence of fining-up, trough cross-bedded gravel units, interpreted as channel-fill deposits, suggests the local development of erosive, channelized flows. The small size and rarity of the trough-shaped units indicates that persistent channelization was uncommon in the fans. Interbedding with diamicton units, interpreted as debris flow deposits, also indicates that fluvial flow within the channels was intermittent. Planar cross-bedded gravel deposits locally reflect the development of foreset beds on transverse bars (Harms *et al.*, 1982). Foreset bedding occurs mainly in channelized gravel successions suggesting an association of transverse bars with deeper channel scours. Interbedded sands indicate intermittent decreases in bedload coarseness, presumably due to variable flow velocities during growth of the bars.

Cross-Stratified Sands (St, Sp)

Cross-stratified sand facies generally consist of moderately to very well-sorted, medium to coarse sands and pebbly sands. Beds exhibit trough cross-laminae (St; Fig. 3-16a) and, more rarely, planar cross-laminae (Sp). Scour and fill structures are commonly associated with these facies. Cross-stratified sand lenses may be over a metre thick and several metres wide and are invariably trough-shaped (Fig. 3-16b). They commonly fine upwards with medium to coarse sands at the base grading up into fine sand or silt facies. They may have scoured erosional bases or gradational lower contacts with underlying sandy gravels. Cross-stratified sandy strata usually are well sorted and interbedded with massive to crudely stratified pebble-gravels or, less commonly, with ripple bedded sands and horizontally laminated muds.

Interpretation

Cross-stratified sand facies are interpreted as the products of bedform migration within fluvial channels. Well developed cross-stratification indicates turbulent fluid flow with traction dominated sedimentation. The development of scour and fill structures reflects erosion of the stream bed under turbulent flow conditions and subsequent infilling by sandy bedforms. Planar cross-bedded sands and pebbly sands (Sp) are interpreted as lower flow regime, straight-crested transverse bar, linguoid bar or sand wave deposits while trough cross-bedded strata (St) are formed by sinuous crested dunes (c.f. Harms et al., 1982).

Deposition of large trough-shaped (channeled) sand units such as those at Amber Mountain (Figs. 3-8 and 3-16b), probably occurred in ice-marginal channels. This interpretation is supported by the position of the sandy units high on the valley side, the presence of lag gravels at the base of the channels, numerous scour and fill structures, and interbedded massive to crudely-stratified gravels interpreted as proximal glaciofluvial deposits.

Ripple Laminated Sands (Sr)

Fine to coarse-grained sands with well developed ripple-laminae are uncommon and occur mainly as interbeds with other sandy facies and with parallel-laminated muds. Sands with ripple structures also occur in small, isolated lenses and thin beds (Fig. 3-16c), overlying coarse openwork gravels where they fill interstitial voids in the upper part of the gravels, or gradationally overlying matrix-filled sandy gravels. Sandy strata with draped laminae and Type-B (depositional-stoss) climbing-ripple cross-laminae (terminology of Ashley *et al.*, 1982) are mainly fine to medium-grained while Type-A (erosional-stoss) ripple cross-laminae are most common in medium to coarse-grained sands. Ripple-laminated sand lenses typically vary from 2 to 10 centimetres in thickness and 20 to 200 centimetres in length and show well preserved stratification. These lenses commonly are near horizontal, trough-shaped, occur in the lee of large clasts (Fig. 3-16c) and are thickest at the upstream end, tapering down-flow. Deformed laminae and load and flame structures may be present. Some thin, irregularly-shaped sand lenses contain convoluted stratification. They are usually interbedded with poorly sorted gravels and have down-valley dips of up to several degrees. Fluid-escape structures, folded laminae, and loaded upper contacts are common (Figs. 3-16d, e) in these lenses.

Interpretation

Sands of this facies are interpreted as lower flow-regime, ripple-bedform deposits with varying proportions of suspension sediments. They likely were deposited as bar-top, scour-fill and crevasse-splay sands. The latter consist in part of traction dominated deposits including medium to coarse sands showing Type-A climbing-ripple cross-laminae with interbedded trough cross-stratified sands and sandy gravels. Also common are fine to medium grained sands dominated by draped ripple laminae and Type-B climbing-ripple cross-laminae. They are interpreted as more distal crevasse-splay deposits formed where currents are reduced but still strong enough for some traction deposition (*c.f.* Bridge, 1984; Collinson, 1986). They are interbedded mainly with parallel-laminated muds reflecting a larger component of suspension sedimentation relative to traction deposition than in the more proximal crevasse-splay deposits.

Small lenses and beds of ripple-bedded sands conformably overlying gravel beds are inferred to be scour-fill and bar-top ripples deposited during waning flow. Interbedding with trough cross-laminated medium to coarse sands, parallel laminated fine sands and thin silt drapes reflects fluctuations in flow during deposition. These well-stratified sands probably were deposited during periods of reduced discharge and decreased sediment load between, or immediately after, major flow events. Finer textures, near horizontal orientation, and interbedding with open work (winnowed) gravels support this interpretation.

Small, isolated, trough-shaped, sand lenses that taper down-flow of large clasts probably were deposited in areas of flow separation in the lee of bed obstructions (Fig. 3-16d). The fine grain-size, good sorting and association with horizontally laminated sediments and mud drapes indicate deposition during lower flows, probably by intermittent streams. Intermittent stream activity is typical of alluvial fans and is common on modern fans in the area. In contrast, deformed and irregularly shaped sand lenses probably were deposited penecontemporaneously with channel gravels. This is supported by the steep and consistent down-channel dip of the lenses and by the presence of interbedded poorly sorted gravels. Soft sediment deformation structures (Figs. 3-16d, e) locally resulted from differential loading and fluid expulsion.

Massive to Horizontally Stratified Sands (Sm, Sh)

Lenses and beds of massive (Sm) or horizontally stratified (Sh) sands are conspicuous but generally uncommon features in the fans (Fig. 3-16f). They are typically well sorted and medium to coarse-grained although fine sands also occur. Beds vary from about 10 to 40 centimetres in thickness and may be traced laterally from a few to tens of metres. Thinly-laminated medium-sand beds occur rarely (Sh). Poorly developed graded bedding and discontinuous, wavy to horizontal beds, usually more than 1 centimetre thick, are sometimes apparent in otherwise massive sands (Sm) due to slight textural differences between beds. Thin gravelly (granular to pebbly) beds and angular, isolated pebbles occur locally within some massive sand beds (Fig. 3-16g). Dish and pillar structures also occur locally. Many beds show iron and manganese staining along undulatory layers that have cross-cutting relationships with primary sedimentary structures.

Interpretation

Massive to weakly bedded sand facies (Sm) are interpreted to be the products of rapid deposition from hyperconcentrated flood flows. The massive structure of these beds is mainly attributed to syndepositional deformation due to loading and pore fluid expulsion, as indicated by the presence of dish and pillar structures. Postdepositional or diagenetic disruption of bedding may also be important. Gradational contacts between coarse and fine strata, rare graded bedding, as well as other features typical of this facies, such as the common occurrence of outsized clasts, lateral discontinuity of bedding, and gradational contacts with gravelly flood flow and debris flow deposits, are typical of sandy hyperconcentrated flood-flow deposits (Smith, 1986, 1987). Due to rapid deposition, strata in sandy hyperconcentrated flood flow deposits are defined only by the relative abundance of coarse grains and do not have distinct grain size differences or sharp, erosional bed contacts (Smith, 1986). Discontinuous horizontal strata may also be produced by long wavelength, low amplitude, dune-like bedforms in streams transporting high suspended sediment loads (Smith 1986 and references therein). Rare beds of thinly laminated, medium sands (facies Sh) may represent upper-flow-regime plane beds (Allen, 1984).

Parallel Laminated (Fl) and Massive (Fm) Clay, Silt and Fine Sand

Facies with modal grain sizes finer than medium to coarse sand are uncommon in the fan successions. They occur as massive to crudely laminated muds (Fm) and parallel laminated fine sand and silt (Fl). Strata are thin (commonly 0.5 to 10 centimetres thick), near horizontal, and laterally traceable for up to a few metres. Units of fine sediments, extending laterally for more than a few metres, were observed only at one locality where they are interbedded with medium to coarse sand and silty diamicton beds. Some silt and clay beds have poorly defined, deformed laminae that drape and infill around large boulders or cobbles protruding upwards from underlying units. These laminae may also conform to the contours of overlying clasts (Fig. 3-16h). Vertical fractures are common in mud rich lithofacies (Fm).

Stratification in parallel laminated silt and fine sand beds (Fl) is generally moderately well developed and horizontal or subhorizontal. Typically, thin beds of this facies gradationally overlie medium to coarse-sand or silty diamicton beds and they are most common at the top of fining-up sequences in sand and gravel lenses. Rarely, they occur in sharp contact with underlying gravel beds. Iron and manganese staining give some beds a stratified appearance. This diagenetic layering often cross-cuts any textural stratification present and is also distinguished by its typically steep inclination and/or highly undulatory nature (Fig. 3-16a).

Organic materials, generally rare in the alluvial fan successions, occur locally in fine-grained lithofacies. For example, wood fragments occur in silt and fine sand facies at the Amber Mountain Section but were observed only in one stratigraphic horizon (Fig. 3-9). Wood fragments typically lie in near-horizontal positions and their long axes commonly are oriented parallel to the paleoslope. Recovered wood fragments from the Amber Mountain Section were identified as Picea (sp.?; R.J. Mott, G.S.C. Wood Report No. 88-23) and yielded infinite radiocarbon dates.

Interpretation

Massive to crudely laminated silts and clays (Fm) are interpreted as ponded water deposits formed by settling of suspended fine sediments. They are analogous to fine sediments deposited in small ponds on modern fans (Fig. 3-6b). Cross-stratification and interbedded sands are rare reflecting the lack of strong currents at the time of deposition. Vertical fractures are interpreted as desiccation cracks and indicate that the ponds were ephemeral, probably consisting mostly of water remaining after flood events. This interpretation is supported by the presence of other structures expected in this depositional environment such as mud drapes over large clasts, thin parallel laminae, and load and deformation structures. Ponding on the fan surface was probably confined mainly to small, isolated lows where water could be trapped, such as scoured areas in small abandoned channels, and consequently deposits of this facies have an aerial extent generally limited to a few metres. The massive structure of many mud beds is probably due mostly to post-depositional or diagenetic disruption of bedding resulting from processes such as desiccation and compaction. Laminae in some mud lenses have apparently been deformed as a result of bed irregularities and by the weight of overlying clasts.

Parallel laminated, fine sands and silts (Fl) are interpreted to have formed by settling of suspended fine-sediment with minor traction deposition. Depending mainly on their lateral extent and their association with other facies, they are interpreted either as overbank deposits or as the final suspension-phase deposits of sediment gravity flows or waning floods. For example, where horizontally laminated muds conformably overlie fining-up sand and gravel sequences, a late temporal association with floods of progressively decreasing energy is suggested. Gradational lower contacts with underlying matrix-filled gravels represent a gradual transition from traction to suspension-settling deposition. Similarly, gradational lower contacts with diamicton facies suggest deposition from turbulent sediment suspensions in the last stage of mud flow events. Beds of this facies that are laterally traceable for several metres or more are interpreted as overbank deposits. They are commonly associated with sandy sediments inferred to be crevasse-splay deposits (*c.f.* facies Sr). Kostaschuk *et al.* (1986) interpreted similar laminated sands on modern alluvial fans in the Rocky Mountains near Banff as overbank deposits.

FACIES ASSOCIATIONS

Three main facies associations are recognized in the alluvial fan successions (Figs. 3-9 and 3-10). Facies Association 1 (FA-1) consists of repetitive successions of gradationally interbedded, clast-supported diamicton facies. These include mainly massive diamicton facies (Dc) and lesser amounts of inversely graded (Dci) and normally graded (Dcn) subfacies. Bed boundaries are typically gradational and distinguished mainly by relative variations in the matrix content of the different beds. Gravel or sand facies in this association are restricted to small lenses or rare thin interbeds. Units of FA-1 generally are from a few to several metres thick. Facies Association 2 (FA-2; Figs. 3-9 and 3-10) is dominated by gravel and sand lenses (mainly facies Gm, Gp, Sr and St) and minor fines (Fl, Fm). These deposits occur in fining-upward successions, commonly above scoured, minor unconformities. Thin beds of clast-supported diamicton (facies Dc) and rare matrix-supported diamicton (facies Dm) are also gradationally, as well as sharply, interbedded with these deposits. Units of FA-2 seldom exceed a few metres in thickness and commonly are less than 1 metre thick. Facies Gm(a) occurs in both FA-1 and FA-2. Facies Association 3 (FA-3, Figs. 3-9 and 3-10) is dominated by massive, matrix-supported diamicton facies (Dm) as well as subfacies Dmi and Dmn. This association is uncommon and is restricted mainly to the upper most parts of the fan sequences (Figs. 3-9 and 3-10).

Facies Association 1 comprises the bulk of the fan successions, in terms of both thickness and frequency, and therefore represents the most significant element in fan construction. Noncohesive debris flows (facies Dc) dominate as do gradational bed contacts suggesting that these sequences are largely aggradational. Erosional surfaces are restricted to the base of small gravel or sand lenses indicating that only relatively minor erosional events occurred during deposition of FA-1. In contrast, erosive contacts are common in FA-2 and gravel and sand lenses are abundant, suggesting that channelized fluvial sedimentation dominated. Facies interpreted as debris flow deposits (*i.e.* Dc and Dm) are uncommon and thin in FA-2. The presence of suspension deposits (Facies Fl and Fm), some with desiccation cracks, at the top of fining-up units indicates that periods of non-deposition punctuated some channel-fill episodes.

Facies associations 1 and 2 are, therefore, interpreted to reflect episodes of debris flow dominated sedimentation (FA-1) and intervening periods of intermittent fluvial sedimentation (FA-2). These latter periods of relative quiescence were interrupted only by minor debris flow events. Shifts from sedimentation of FA-1 to FA-2 may have been triggered by aggradation of noncohesive debris flows in FA-1, eventually resulting in a shift of the depositional focus towards lower, adjacent parts of the fan. This would allow for periods of non-deposition. Subsequently, thin beds and lenses of gravel and sand in FA-2 may have been intermittently deposited as overbank sediments or in small channels formed by surface runoff. Repetitive successions of FA-1 and FA-2 (Figs. 3-9 and 3-10), thus, may represent cycles of bed aggradation, channel abandonment and eventual reoccupation.

The occurrence of facies Gm(a) and Sm in association with both FA-1 and FA-2 (Figs. 3-9 and 3-10), may reflect the observation that gravelly and sandy hyperconcentrated flood flows are transitional between stream flow and debris flow deposits (*C.f.* Smith, 1987). They thus occur, with nearly equal frequency, in settings dominated by either fluvial or debris flow deposits.

The presence of Facies Association 3, dominated by cohesive debris flow deposits (Facies Dm and subfacies Dmn and Dmi), mainly at the top of the fan successions indicates that a substantial change in depositional environment occurred near the end of fan aggradation. This is interpreted to directly or indirectly reflect changes induced by the onset of glaciation. Direct changes may include the transport of debris, more conducive to cohesive debris flow production, into the fan drainage area by glaciers. Indirect effects include climatic changes related to glaciation that, for example, may have resulted in a reduced vegetation cover or increased permafrost in the fan drainage basin. Such changes may have produced a new debris source or caused a shift in triggering mechanism (for example from storm-triggered to slump triggered), allowing for the propagation of cohesive rather than noncohesive debris flows.

DISCUSSION

Depositional Processes in Gravel-Dominated Alluvial Fans

In alluvial fan environments, sediment and water is funneled from topographically and geomorphologically diverse drainage areas onto fan surfaces in response to a wide spectrum of climatic and geologic events. Runoff characteristics (*e.g.* volume and rate), origin and amount of sediment load, and mechanisms of sediment supply and transport vary from event to event, and may even change during single events (Bull, 1972). Facies analysis of the paleoalluvial fan successions described indicates that, although coarse-grained debris flow deposits dominate, their sedimentary characteristics reflect this diversity in processes of sediment transport and deposition. Sedimentologic variations within these deposits commonly suggest that continual changes in flow type occurred both from event to event and within individual events.

A wide variety of flow mechanisms are known to deposit coarse gravelly sediments, varying along two more or less continuous sequences (Fig. 3-17) between normal stream flow and two types of sediment gravity flows: grain flows and cohesive debris flows (Lowe, 1979, 1982; Shultz, 1984; Smith, 1986). Although end member types can readily be distinguished, there are many more examples of intermediate varieties. This range of depositional regimes on the paleofans is discussed by considering three constitutive components of the flows: granular materials, fine sediments (particularly clay) and water (air generally being a minor component in most flow types; Pierson and Costa, 1987). The first two components can be directly measured from the preserved deposits whereas the third must be inferred, mainly from sedimentologic data. On the conceptual ternary classification shown in Figure 3-17, end member types of subaerial gravelly flows are approached by increasing either water, clay or clast content. During sedimentation, when most preserved sedimentologic characteristics form, these three components have critical effects on flow rheology, which is determined mainly by turbulence, clast interactions and the plastic versus fluidal behavior of the flow (Postma, 1986).

End-Member Flow Regimes

Cohesive debris flows (in which clasts are supported by a cohesive matrix), also referred to as mudflows, true debris flows or viscous slurry flows (Middleton and Hampton, 1976; Lowe, 1979, 1982; Pierson and Costa, 1987), compose the most clay-rich, end member in the classification of flow types shown in Figure 3-17. These flows have sufficient yield strength to support gravel-sized clasts in suspension and viscous forces control flow behavior. Inferred cohesive debris flow deposits in the Jasper fan successions, are invariably massive, ungraded, matrix-supported diamictons (facies Dmm). Flow occurs as a rigid plug riding above a zone of continuous shear with high pore water pressures in which a basal inversely graded layer (subfacies Dmi) may develop (Postma, 1986). In flows of lower strength, rare coarse-tail normal grading (Dmn) may develop due to settling of clasts whose individual weights exceed the yield strength of the plug (*e.g.* Shultz, 1984; Smith, 1986).

Grain flows, forming another end member flow type characterized by high clast concentrations (Fig. 3-17), are rare in the Jasper alluvial fan successions. Particle support in true grain flows (Lowe, 1982) is only by dispersive pressure with deposition occurring by frictional freezing. Grain flows occur mainly on steep slopes and are thin because dispersive pressure at the base of the flows is generally insufficient to support thick sediment columns (Middleton and Hampton, 1976; Lowe, 1976, 1982). However, in cohesionless grain dispersions with mixed sediment sizes (density-modified grain flows), buoyant lift of the sediment-water mixture can support larger particles and allow for development of thicker gravelly flows (Lowe, 1976, 1982; Walker, 1977). Possible grain flow deposits on modern fans in the region are restricted to upper fan regions, where high velocity fluid flows first issue out of the steep fan head channels, suggesting that gradients lower down on the fans are insufficient to maintain grain flows.

Clast-supported, coarse-gravel lithofacies (Gm, Gp, Gt) with well developed hydrodynamic structures are the result of the most fluid, sediment-poor, end member in this sequence of flow types (Fig. 3-17). These flows are characterized by turbulence and suspended sediment concentrations that are not sufficiently high to play a significant role in sediment support . Stratification and imbrication are a result of traction current sedimentation. Most deposition occurs from confined (channelized) flows following peak flood discharges capable of transporting large clasts, and to a lesser extent from unconfined (sheetflood) flows. Massive or horizontal to low angle, crude bedding (facies Gm) in most channel gravel deposits indicates deposition from diffuse gravel sheets, longitudinal bars or diagonal bars.

Transitional Flow Regimes

Flows with characteristics intermediate between these three end member types (Fig. 3-17) have been described by various authors as gravelly debris flows (Larsen and Steel, 1978; Burgisser, 1984), density-modified grain flows (Lowe, 1976, 1982), pseudoplastic debris flows (Shultz, 1984), hyperconcentrated flood flows (Smith, 1986), clast-rich debris flows (Waresback and Turbeville, 1990) and noncohesive debris flows (Scott *et al.*, 1992). Similarly, in Postma's (1986) theoretical classification of sediment gravity flows, based on flow characteristics at the time of deposition, three intermediate types of debris flows with high sediment concentrations are distinguished: laminar fluidal (cohesionless) flows, turbulent fluidal flows and turbulent cohesive (plastic) flows. Using this classification, most clast-supported diamicton facies (Dc) in the fan successions are interpreted as the deposits of laminar, cohesionless debris flows, since turbulence is unlikely in the final rheological phase of these high concentration sediment flows. Postma (1986) also noted that turbulence in flows with high sediment concentrations would be rare except possibly in some subaqueous pyroclastic flows or some subaerial cohesive debris flows that should produce large-scale eddy structures. Beds of facies Dmn with erosional lower contacts may have formed from the latter type of flow. The rheology of some poorly sorted gravelly flows, interpreted here as hyperconcentrated flood flows (subfacies Gm-a), may at the time of deposition also approach that of turbulent, high concentration, cohesionless flows (Fig. 3-17).

Clast supported diamicton facies (Dc) that dominate the fan successions are interpreted here as noncohesive or pseudoplastic debris flows. Lift in these flows is provided in various degrees by buoyancy, escaping pore fluids, dispersive pressure and turbulence depending in part on the relative concentrations of clasts, fines and water. Most clast-supported beds are ungraded indicating that dispersive pressure resulting from grain collisions was low. Postma et al. (1988) suggested that inertia flows resulting in ungraded deposits of this type are different than traction carpet flows and may range from cohesionless flows to cohesive slurries. Cohesionless sediment gravity flows are characterized by laminar or pseudo-laminar flow in high-density, fully sheared, inertia-flow layers (Lowe, 1979; Postma, 1986). Flows of this type are similar to the inertial slurry flows of Pierson and Costa (1987) in which estimated flow velocities of approximately 2 m/sec or more are required for inertial mechanisms to dominate over viscous forces. Similar velocities in flows of this type have been reported from Cathedral Mountain debris flows in the southern Canadian Rocky Mountains (Jackson et al., 1989). Eyewitness accounts of debris flow events at Cathedral Mountain that crossed the Trans-Canada Highway, indicate flow velocities of up several metres per second although much slower movements were reported in the lower deposition area; calculated velocities in the fan head channel ranged from 3.1 to 5.5 m/sec (Jackson et al., 1989).

In some beds of facies Dc, particularly those dominated by coarse grains, dispersive pressures were sufficiently high for inverse grading to develop (subfacies Dci). As a result of turbulence in these flows, the concentration of large clasts, may increase towards the base resulting in enhanced dispersive pressures there and the development of an inversely graded, basal layer. Freezing of such a flow may occur nearly instantaneously when shear stress drops below the yield stress and may be followed by a suspension sedimentation stage resulting in an inversely to normally graded deposit (Lowe, 1982). Turbulence in noncohesive flows may also develop stratification that rarely may be preserved as in subfacies Dcs.

Gravels in the paleofan successions inferred to be hyperconcentrated flood flow deposits typically have a massive structure or crude bedding, bimodal clast-fabrics, distribution-normal grading and a polymodal matrix (subfacies Gm-a and possibly also Dcn). Lithofacies with similar sedimentologic characteristics have been described from numerous other alluvial fan successions (Wasson, 1977; Nemec and Steel, 1984; Smith, 1986) and analogous settings in volcanic terrains (Beverage and Culbertson, 1964; Pierson, 1981; Scott, 1986; Major and Voight, 1986; Walton and Palmer, 1988; Palmer and Walton, 1990). These high discharge, cohesionless, partially turbulent flows have unusually high sediment loads and depositional characteristics intermediate between fully turbulent, traction-dominated stream flows and laminar debris flows that deposit sediment *en masse* (Smith, 1986). Hyperconcentrated flows have a measurable but low yield strength (Pierson and Costa, 1987). Turbulence is dampened by high sediment concentrations (>30 % by weight; Beverage and Culbertson, 1964; Weirich, 1989) but remains the principal sediment-support mechanism, although dispersive pressure and buoyancy are also important. Sedimentation in these flows is characterized by grain-by-grain aggradation from both suspension and traction, typically producing bimodal fabrics and normal grading. Laminar shear may selectively rearrange small clasts in the traction load (Fig. 3-15) and may also produce horizontal pseudolamination but dispersive stresses seldom are sufficiently high to develop inverse grading (Postma *et al.*, 1988).

Although turbulence and to a lesser degree dispersive pressure are the primary grain support mechanisms of hyperconcentrated flood flows (Smith, 1986), at the time of deposition a highly concentrated traction carpet undergoing late-stage laminar flow may develop in the lower part of the flow (Postma *et al.*, 1988; Postma, 1986; Smith, 1986). Although Smith (1986, p.4) stated that hyperconcentrated flood flows are not sediment gravity flows, he noted that they produce deposits similar to coarse-grained, high-density, subaqueous, turbulent flows and suggested that both flow types have similar mechanisms of sediment support, transport and deposition. The density-modified grain flows of Lowe (1976, 1982) are similar to hyperconcentrated flood flows in that they also form clast-supported, pebble to cobble gravel deposits with a poorly sorted matrix.

Although noncohesive debris flow and hyperconcentrated flood flow deposits may not always be distinguishable, the presence of normal or inverse grading and detailed fabric data can locally be used to differentiate them. Hyperconcentrated flood flow deposits frequently, but not always, show distribution normal grading and the long-axis orientations of clasts have bimodal distributions that are a function of grain size (Smith, 1986, 1987). Large clasts transported in the flows by traction are oriented with long-axes perpendicular to flow and smaller clasts, transported above the bed, are oriented with long-axes parallel to flow (Fig. 3-15). In contrast, noncohesive debris flows may contain inversely graded beds, particularly at the base of coarse (cobble to boulder-sized) gravelly flows, that may be overlain by suspension sedimentation units characterized by coarse-tail normal grading (Lowe, 1982; Postma, 1986). In addition, inverse to normally graded conglomerates typically contain clasts with a preferred orientation of long-axes parallel to flow, often with an upstream dip (Walker, 1975).

A continuous spectrum of flow and deposit types probably exists between gravelly, noncohesive debris flows and hyperconcentrated flood flows and both may be part of a larger continuum (Lowe, 1982; Scott et al., 1992). The relative importance of dispersive pressure and turbulence as grain-support mechanisms during sedimentation is probably the main control on the final deposit type. Normal grading in some clast-supported diamicton beds (Dcn) and the local presence of stratification, clast clusters and scoured surfaces (subfacies Dcs) indicate that some cohesionless debris flows may be transitional with more turbulent flows such as hyperconcentrated floods.

Flow Transformation

Lateral changes in grain-size distribution and sedimentologic character, especially common within gravelly diamicton beds (facies Dc, subfacies Dci, Dcs and Dcn), is inferred to reflect changes in flow character resulting from variations in the relative proportions of water and sediment. Such rheologic changes in debris flows are referred to as flow transformations (Fisher, 1983). Relative increases in sediment concentration may occur as a result of infiltration of water into the substrate or as a result of erosion of sediment from the bed and incorporation into the flow. Conversely, progressive dilution of the flow can occur gradually as a result of heavy rainfall and surface runoff or, more suddenly, as a result of the addition of water at channel confluences. Gradations between clast-supported and matrix-supported facies in the fan deposits, for example, may reflect these temporal and spatial variations in water and sediment concentrations. Transitions between cohesive and noncohesive flow during single flow events have been reported by numerous authors (Johnson, 1970; Lowe, 1979; Middleton and Hampton, 1976; Nardin et. al., 1979; Smith, 1986; Scott et al., 1992). Flow transformations of hyperconcentrated flood flows to and from noncohesive debris flows appear to be the most common whereas cohesive debris flows do not transform as readily (Smith, 1986; Scott et al., 1992). In the Jasper fans, diamicton beds grading-up into massive gravels, pebbly sands and finally into silts (Fig. 3-9) may reflect flow transformations from debris flows to hyperconcentrated flows and lastly stream flows and suspension deposits. A similar transformation was recorded in a modern flow by Weirich (1989) both by a progressive decrease in peak sediment concentration from >65% to <40% and finally to <10% and by a facies transition from structureless, debris flow diamicton to poorly sorted, weakly stratified, hyperconcentrated flow deposits and lastly to well stratified and well sorted fluvial deposits.

Transformations between different types of flows, resulting from rapid changes in sediment or water input into single channels, are probably more common in alluvial fans than other depositional environments since fan channels commonly focus converging flows from numerous tributary streams, each reflecting potentially distinct geomorphic, geologic and hydrologic conditions in different parts of the drainage basin. Hyperconcentrated flood flows, for example, may develop by rapid dilution of debris flows with runoff waters from storm or snowmelt events, or they may transform to cohesionless debris flows as a result of sudden increases in sediment load in streams, usually caused by slope failures or rapid increases in bed erosion during flood discharge (Smith, 1986). Transitions from flood flow to debris flow may also be common due to infiltration, particularly on coarse-grained fans. Flow transformation in debris flows, due to progressive dilution of the flows as they encounter water in pre-existing stream channels, is unlikely on arid region alluvial fans (Pierson and Scott, 1985; Smith, 1986) but may be more common in temperate region fans where water is present in fan channels for much of the year and melting of high altitude snow may contribute significantly to spring and summer discharges. These factors may help explain frequent transitions between different debris flow facies and the common development of intermediate flow types in the temperate region fans described here.

Although uncommon, cohesive debris flows may develop from cohesionless flows and there may also be a continuous flow spectrum between these types (Lowe, 1982; Postma, 1986). For example, normal grading in subfacies Dmn may be inherited from an early turbulent phase, prior to transformation from noncohesive to cohesive flow. Likewise, incorporation of clay into the matrix of some clast-supported diamicton beds of facies Dc, may have resulted in transformation of noncohesive to cohesive debris flows. Lowe (1982) observed that cohesive debris flows vary from flows in which the larger clasts are totally supported by the buoyancy and cohesiveness of the clay-water matrix to those where clasts, comprising up to 95 percent of the flow by volume, remain more or less in contact. Such flows are characterized by clay, even if minor, in the fluid phase, reduced effective normal stress and low interlocking of clasts resulting in reduced friction.

Fluvial Depositional Environments

Although fluvial sediments are much less dominant in the paleofan successions than debris flow deposits, they probably represent deposition during relatively long periods of time. On modern fans in the study area, nearly continuous stream flow occurs in main channels throughout the spring and early summer months but debris flow events are relatively rare. Stream flows have low sediment loads and their sedimentologic importance is mainly that of slight winnowing of debris flow tops and deposition of trough-shaped, open-work, channel lag gravels, one or two clasts thick, and small sand and gravel lenses (Figs. 3-14d, 3-16). The common upwards gradation of massive matrix-filled gravel beds into open work beds reflects winnowing of the matrix by sediment depleted flows or intermittent stream activity. Winnowed gravels can readily be observed on modern alluvial fan channels in the region (Fig. 3-5c).

Fluvial sedimentation in the paleofans, particularly in the Front Ranges, probably occurred mainly in shallow streams as indicated by the dominance of thin, subhorizontal bedding in fluvial deposits interpreted as diffuse gravel sheets and longitudinal or diagonal bar gravels. The abundance of these low relief bedforms reflects shallow streams, coarse bedloads and rapid deposition, conditions typical of alluvial fan environments. High relief in the drainage basin and high flows, following unusual storm and snow-melt induced discharge events, allow for the development of coarse gravel bedloads. Sedimentation occurs rapidly below the intersection point where slopes sharply decrease and flows spread out and thin. Shallow channels with thin gravel deposits dominate modern fans in the area below their intersection points (Fig. 3-4). The rarity of planar and trough cross-bedded gravel facies (Gp, Gt) in the paleofan successions suggests that conditions generally were not suitable for the development of transverse bars or gravel dunes, probably also due to the shallow geometry of the channels. Cross-bedded gravels and sands and scour-and-fill structures are restricted mainly to the upper parts of the paleofans at a few sites (*e.g.* Amber Mountain, Fig. 3-16b), where they are transitional with glaciofluvial braided stream gravels. Persistent channelization was otherwise uncommon in the fans as indicated by the rarity of channel gravels and their common interbedding with debris flow deposits and quiet water sediments.

Unconfined flows over permeable sediments at the mouths of some fan channels may have locally resulted in the deposition of massive, angular, gravels interpreted as sieve deposits. They probably formed when rapid loss of water and matrix-fines, due to infiltration, resulted in a dense dispersion of grains with increasing frictional resistance leading to rapid deposition. The limited aerial extent of these deposits in both modern and paleofan successions in the study area indicates that conditions suitable for their development are uncommon.

Progressively decreasing discharge in the fan channels is indicated by massive and parallel-laminated clays, silts and fine sands (facies Fm and Fl) that cap some fining-up sand and gravel lenses (Fig. 3-9). Deposition of the fine sediments probably occurred in the final waning phases of the flows. Interbedding of cross-stratified sand facies and fine gravels (Fig. 3-16a) reflects variations in current strength. Debris flow and flood flow deposits also locally fine-up through parallel laminated sands and silt (Fl) to massive muds (Fm), reflecting localized suspension sedimentation at the termination of the main debris and flood flow events. Desiccation cracks in the capping mud drapes indicate subaerial exposure and attest to the intermittent nature of the flows.

Non-channelized fluvial sedimentation on the paleofans apparently was restricted mainly to minor back levee deposits. Laterally extensive beds of sand and gravel at Amber Mountain (Fig. 3-8), for example, probably were deposited behind channel levees during flood events. This interpretation is supported by interbedding of facies interpreted as crevasse-splay and over bank deposits and the lack of associated channel gravels. Sedimentation was not restricted behind the channel levees, thus allowing for the development of beds several tens of metres wide.

Controls on Sedimentation Patterns

Intrinsic versus Independent Factors

Vertical changes in the distribution of lithofacies within the fan successions, recognizable at individual sites, can generally be attributed to intrinsic factors such as shifts in the locations of feeder channels, but independent factors also appear to be important in the latter stages of fan development as indicated by similar lithofacies changes near the top of the fan successions at a number of sites in the region. For example, at the Roche Miette fan, an increase in abundance of matrix supported diamicton beds at the top of the section (Fig. 3-10) reflects superposition of cohesive debris flows over fluvial channel and noncohesive debris flow deposits. This may be a result of climatic changes such as increased precipitation or other effects related to impending glaciation, or may be due to more intrinsic factors such as a lateral shift or down-fan migration of the fan head trench. The former interpretation is preferred because of the association of the uppermost fan deposits with glaciofluvial gravels at this and other sites in the region.

A change on the north side of the Amber Mountain fan from gravelly debris flow dominated sedimentation to fluvially dominated sedimentation for a short interval in the middle of the fan succession, is indicated by a unit of muddy sediments (Fl, Fm) and interbedded sands and gravels that can be traced along the entire longitudinal profile (Fig. 3-8, top). The sediments are thickest at the downstream end of a bedrock high and they gradually thin down-fan towards the southwest. Ripple bedded sands (Sr), parallel laminated fine-grained sediments (Fl) and muddy diamicton deposits (Dm)are common in this unit; they reflect crevasse-splay, overbank and debris flow sedimentation, respectively. The presence of these low energy sediments rather than coarse noncohesive debris flow deposits may be due to either a lateral shift in the depositional focus of the main fan feeder channel or a change in climatic or other controls resulting in an interval of relative quiescence before renewed deposition of higher energy deposits. A shift in the focus of sedimentation is the favored interpretation for this unit because of the lack of lateral continuity of these facies across the fan (Figs. 3-8a, b). This shift in deposition towards the south probably was caused by a bedrock high on the up-stream end of the north side of the fan (Fig. 3-8a) that initially restricted debris flow sedimentation below that point. During that time, deposition of finer sandy facies occurred dominantly as a result of over topping of the main channel levee to the south or temporary channel avulsion during high flood flows. Fan aggradation on the south side of the fan, to a level above the influence of the bedrock high, eventually allowed for lateral migration of the main feeder channel back to the north and the return to gravelly debris flow dominated sedimentation. A climatic influence on the development of this extensive unit of relatively low energy deposits cannot, however, be entirely discounted by its restricted lateral extent, as it may have been eroded from the south side of the fan by subsequent flows. This, however, is unlikely as sandy and even silty sediments are commonly found preserved beneath coarse debris flow deposits and it is therefore unlikely that all evidence of such an extensive unit could have been removed from the south side of the fan.

The lack of pronounced erosional surfaces indicating channelization and the dominance of gradational bed contacts in most gravel and diamicton facies suggests that deposition occurred primarily below the paleointersection point and not in erosive, channelized parts of the upper fan. Modern fan head trenches in the area are erosional and little or no sedimentation occurs within them and it is therefore unlikely that deposition in fan head channels, above the paleofan intersection points, can account for the thick successions of deposits exposed. The presence of coarse gravelly debris flow deposits throughout the length of the fan successions also suggests that they were deposited in the main body of the fan (medial and distal portions) and not only in the upper fan reaches. However, all of the described fan successions are exposed along modern fan head trenches indicating that the paleofans were graded to a substantially higher (~100 m) base level than the modern valleys (the slope of modern fan head channels being similar to the dip of the paleofan bedding). Thus, intersection points on the paleofans must have been significantly higher and closer to the valley side than at present.

Many models of alluvial fan sedimentation depict coarse-grained, thick, debris flow deposits dominating upper fan reaches and fluvial deposits, of progressively decreasing grain-size and bed thickness, in more distal, down-fan reaches (*e.g.* Reineck and Singh, 1980; Mack and Rasmussen, 1984; Rust and Koster, 1984; Hooke, 1987). Obvious factors controlling these sedimentation patterns typically include down-fan decreases in slope, channel depth and proximity to the sediment source. However, an analysis of lateral facies relationships at the Amber Mountain and Roche Miette sites (Figs. 3-8 to 3-10) reveals a general lack of such consistent trends. Lateral changes in the frequency of debris flow deposits relative to fluvial facies are not readily apparent and the thickness of diamicton beds and modal grain-size of coarse gravel facies do not systematically change down-fan. One possible explanation for this may be that the Jasper paleofans were too steep and short to allow for significant down-fan variations in sedimentation pattern. The dominance of gravelly debris flow deposits in both proximal and more distal parts of the fan successions, particularly at Amber Mountain where the fan sediments are exposed over much of the longitudinal profile of the paleofan, supports this interpretation. A similar conclusion was recently reached by Blair and McPherson (1992, 1993) who questioned the Trollheim fan model by showing that debris flow deposits dominate all parts (rather than just the proximal reaches) of fans at the type locality. Whipple and Dunne (1992) also demonstrated that steep fans built by debris flows with high sediment concentrations may not exhibit pronounced proximal-distal variations. Nevertheless, it is possible that some down-fan changes in the Jasper paleofans would have been detected if the more distal, eroded reaches of the paleofans could have been observed. This is also suggested by the presence of significant down-fan changes only in the most distal portions of Quaternary alluvial fans in other areas (*e.g.* Harvey, 1984). However, as discussed below, bedrock controls probably have a stronger influence on the relative distribution of debris flow and fluvial facies than paleogeographic factors (*i.e.* upper fan versus distal fan position).

Bedrock Controls: Contrast between Main Ranges and Front Ranges

An obvious contrast between Main and Front Range fan deposits results from differences in the relative abundance of fluvial and debris flow facies. In Main Range fans, very poorly sorted, coarse gravelly, debris flow deposits (facies Dc) dominate and fluvial gravel facies are relatively uncommon. In the Front Ranges, shallow gravelly stream and flood flow deposits (facies Gm and Gm[a]) are commonly interbedded with debris flow deposits. Similarly, sandy facies occur as thin trough-shaped lenses throughout the Front Range fan successions. However, in the Main Ranges they are relatively rare except at Amber Mountain where they are confined mainly to the one stratigraphic interval discussed above (Fig. 3-8).

In the Main Ranges, noncohesive debris flow beds (facies Dc) are generally fewer and of greater thickness than in the Front Range fans, although individual beds are often difficult to distinguish due to gradational bed contacts. The average thickness of noncohesive debris flow beds in the Front Ranges is 0.65 metres (mean of 68 beds ranging from 0.2 to 1.5 metres thick) compared to 1.9 metres in the Main Ranges (mean of 22 beds ranging from 0.3 to 4.8 metres thick). Another distinction is due to differences in clast size. Noncohesive debris flow deposits in the Main Range fan successions are dominated by cobble to boulder-sized deposits whereas pebble to cobble-sized materials dominate in the Front Ranges (Figs. 3-7, 3-9 and 3-10). These deposits also generally have fewer matrix fines in the Front Ranges than in the Main Ranges. Similarly, massive diamicton beds interpreted as cohesive debris flow deposits, relatively rare in the fan successions, generally have higher clay contents in the Main Ranges than in the Front Ranges. These data indicate that debris flows in the Main Ranges were more poorly sorted and larger than those in the Front Ranges. Assuming approximately equivalent sedimentation periods in both regions, the higher number of diamicton beds of facies Dc in the Front Range fans also suggests that noncohesive debris flows occurred more frequently there than in the Main Ranges. In addition, the greater thickness and relatively coarse grain-size of debris flow deposits in the Main Range fans suggests that debris flow events there generally were of higher energy and transported more sediment than in the Front Ranges.

These differences in frequency, magnitude, competence and flow type are considered to largely reflect differences in bedrock geology. Clastic source rocks in the Main Ranges, consisting of quartzose conglomerate and sandstone interlayered with limestone, shale and siltstone, are considered to more conducive to the formation of noncohesive debris flow deposits containing numerous large detrital clasts than are the massive, Devonian Palliser limestone beds that dominate in the Front Ranges. Drainage basins dominated by resistant rock types, such as quartzitic rocks in the Main Ranges, typically yield large quantities of coarse bedload to stream channels (Kochel, 1988). The relatively small modal grain size of limestone and shale clasts in the Front Range deposits is also probably due to numerous, argillaceous, strata thinly interbedded within the Fairholme Group limestones. Similarly, the relatively high frequency of debris flow events in the Front Ranges may be explained by more complex interbedding and deformation of the source rocks making them more susceptible to small, frequent, debris slides. More thickly bedded and gentler dipping rocks in the Main Ranges, may be mobilized only during relatively rare triggering events, thus accounting for their lower frequency and possibly also their higher energy. Finally, compared to Main Range drainage basins of similar size, Front Range basins are long and narrow due to steep imbricate-thrusting. This may contribute to more uniform discharges during flood events in the Front Ranges and may account for the typically thinner beds and smaller maximum grain-size of transported materials. Conversely, basin width/length ratios, closer to unity in the Main Ranges, probably result in more quickly focused run off and flashy discharge. This may lead to the larger magnitude debris flow events that characterize the Main Range fans.

Kostaschuk *et al.* (1986) found a similar relationship between bedrock geology and fan type. They observed that debris flow dominated fans formed at the base of small, steep drainage basins whereas larger, less steep basins produced fans dominated by fluvial processes. Although the Amber Mountain and Roche Miette fans are similar in size, the greater length to width ratio of the Roche Miette drainage basin accounts for the less steep basin slopes (Table 3-1) and the greater abundance of fluvial deposits relative to debris flow deposits.

Harvey (1984) also found that depositional processes on alluvial fans are influenced mainly by drainage basin size, basin relief and local geology, in addition to climate. Alluvial fans with small steep drainage basins in massive sedimentary rocks, yielding both coarse clasts and adequate fine sediments, favour debris flow deposition, whereas larger, less steep basins on rocks that yield little fine sediment favour fluvial deposition. The Jasper alluvial fan successions occur well within the range of debris flow fans as defined by Harvey (1984) from a study of 31 Quaternary alluvial fans in Spain. Debris flow fans there were distinguished from fluvial fans by the presence of at least 20% debris flow deposits. Debris flow fans were also characterized by high basin relief (H_d) and steep fan slope (S_f) in relation to planimetric basin drainage area (A_d) as defined by the expressions:

$$H_d = 300\,A_d^{0.69} \quad \text{and} \quad S_f = 0.079A_d^{-0.208}$$

where basin relief is in metres and drainage area is in square kilometres. Using these discriminant equations, calculated boundary relief and slope values for the Amber Mountain and Roche Miette fans are 810 metres, 6.1° and 732 metres, 5.9°, respectively. By comparing these calculated values with actual measured values (Table 3-1) it can be seen that the Amber Mountain and Roche Miette paleofans fall well within the range of debris flow fans, having substantially higher basin relief and steeper slopes than the fluvial fans studied by Harvey (1984). This conclusion is supported by the dominance of facies interpreted as debris flow deposits in the Jasper area fans (> 50% at all sites).

In similar studies of a large number of alluvial fans in the southern Canadian Rocky Mountains, Jackson (1987) and Jackson *et al.* (1987) found that fans constructed, at least in part, by debris flows had slope angles of 4° or greater, whereas lower slopes characterized fans of strictly fluvial origin. Similarly, debris flow fans were characterized by drainage basins predominantly smaller than 10-12 km^2 and with relief's of more than 500 m. The drainage basins of the Jasper fans are about one third this size and have more than twice the relief (Table 3-1). Likewise, Kostaschuk et al. (1986) and Jackson *et al.* (1987) found that plots of fan slope versus Melton's (1965) ruggedness number (R), defined according to the equation:

$$R = H_d A_d^{-0.5}$$

with H_d and A_d as defined above, are useful for discriminating fluvial and debris flow fans. Most fans of fluvial origin have R values of less than 0.3 and slopes of less than 2.5° (Jackson *et al.*, 1987). By comparison, the Jasper paleofans have R values of 0.5 to 0.6 and slopes of 3° to 14° (Table 3-1), well within the range of the debris flow fans.

Climatic variations within the region apparently have less effect on sedimentation patterns than do bedrock controls. The main climatic difference between the Main and Front Ranges is higher precipitation in the former, due to a rain shadow effect in the lee of the topographically high Main Ranges (prevailing winds are westerly resulting in an increasing aridity gradient to the east). However, higher precipitation in the Main Ranges suggests that the abundance of fluvial deposits should be also higher but the opposite is observed. Common interbedding of fluvial and debris flow facies in the Front Range fans indicates that fluvial sedimentation and reworking of older debris flow deposits occurred frequently there. Similarly, if higher precipitation is responsible for triggering debris flow events, then the relative frequency of such events should also be greater in the Main Ranges but this pattern also is not observed. In addition, a change in relative abundance of debris flow and fluvial facies from west to east within each range would be expected if precipitation is a major factor controlling fan sedimentation, but no such change is observed (Fig. 3-7). Finally, the fan deposits at Roche Bonhomme, at the west side of the Front Ranges, contrast just as sharply with the nearby Main Range fan deposits at Amber Mountain, as do Front Range fans much further east, further suggesting that it is unlikely that climatic differences between the two regions are sufficient to explain the observed variations in facies abundance.

Paleoenvironment

The similarity of the Jasper paleofan deposits to modern fan sediments in the area suggests that they accumulated under similar, cool-temperate, environmental conditions. Important environmental factors in the region, typical of many high elevation mountainous areas, include moderately high precipitation (>400 mm/year), a large proportion of which is snow, and low tree lines in the drainage basins. Perennial snow banks occur locally. Periglacial and nivational processes, physical weathering of exposed rock and lack of vegetation in the high alpine areas that dominate much of the modern drainage basins, allow for the development of an unstable, surficial cover. Frost shattered, coarse-grained scree with little mud dominate these sediments. The presence of unstable surficial deposits combined with the influence of relatively high precipitation and snowmelt, probably resulted in the frequent, gravelly, sediment rich debris flows that characterize the Jasper paleofans. The rare presence of spruce tree fragments in the Jasper fan successions suggests at least some tree cover, but their typical absence implies that the paleodrainage basins were largely unvegetated, not unlike the modern basins.

Mud poor, clast-supported fan gravels, similar to those that dominate the Jasper paleofan successions occur in the South Orkney Islands (Wells, 1984). They are dominated by polymodal, unchannelled conglomerates; about half, with a clay-poor matrix and unorganized fabric, interpreted as debris-flow deposits and half clay free, weakly imbricated, locally well sorted and interpreted as sheetflood deposits. Also similar to the Jasper fan deposits are steep beds, large clasts and the rare presence of tree logs. These were inferred to indicate an alpine setting with a low tree-line and the associated cool, wet, environmental conditions were also attributed to rapid production of an unstable debris mantle promoting debris flow activity (Wells, 1984).

The gravel-dominated debris flow and flood flow events that built the Jasper fans, were probably triggered by storms or sudden snowmelt events. This interpretation is corroborated by recent investigations in southeastern British Columbia that indicate that most debris flows and landslides occur during rain-on-snow or peak snowmelt events (Jordan and Toews, 1993). The most critical time appears to be in the spring during the late snow-melt period when the ground in the drainage basin is saturated and partially snow-free. Snow may also act as a lubricant during debris flow events (Wells, 1984).

Comparison with other Depositional Environments

The Jasper fan successions are similar in many respects to deposits of some proximal glacial and glaciofluvial environments but can be distinguished from them in several ways. Due to glacial communition, silt and clay proportions are high in most glacial and paraglacial environments and therefore cohesive debris flows dominate over noncohesive flows. The resulting deposits are commonly massive, matrix-supported diamicton beds interstratified with sorted sediments deposited between debris flow events. In contrast, the temperate fan deposits described here are characterized by crudely stratified, clast-supported lithofacies deposited mainly by noncohesive debris flows. In addition, glacigenic sequences contain a high proportion of erratic and glacially abraded clasts in comparison with the paleofan deposits where they are rare or absent.

The Jasper alluvial fan deposits are distinguished from glaciofluvial successions by less well developed and fewer fluvial channel facies and by more frequent interbedding of debris flow deposits in alluvial fans. They are also distinguished from glaciofluvial and glacigenic debris flow deposits by the consistent dip of strata towards the valley centre and by lithologic data which indicate provenance from restricted tributary drainage basins on the valley sides (Levson and Rutter, in press). In ancient fan successions where tectonic disruption limits the usefulness of lithologic or paleoflow data, they may be distinguished from glaciofluvial deposits by the dominance of gravelly diamicton beds exhibiting relatively poor sorting, weak or no stratification, coarse grain-size and high clast angularity.

The fan deposits described in this chapter occur in areas of relatively high precipitation, compared to arid region fans. Sedimentation in arid fans is dominated by cohesive debris flows as well as sheetfloods or overland flows that form (by deposition or winnowing, respectively) geometrically distinct, tabular gravel beds with high length to width ratios; channel-fill and sieve deposits are relatively uncommon or absent (Bull, 1972; Blair and McPherson, 1992, 1993). In contrast, the Jasper fan successions are characterized by a higher proportion of gravel and gravelly diamicton lithofacies as well as a preponderance of small channelized deposits, particularly in Front Range fans, that show a discontinuous lens-like, rather than continuous sheet-like, geometry. This suggests that sedimentation in temperate, compared to arid, alluvial fans may be characterized by higher proportions of noncohesive debris flow, hyperconcentrated flood flow and channelized stream flow deposits and fewer cohesive debris flow and sheetflood deposits. These distinct sedimentation patterns in temperate regions probably result from higher and more frequent precipitation, less clay production (due to the dominance of physical over chemical weathering processes) and a greater influence of snowmelt events than in arid regions.

CONCLUSIONS

Catastrophic flow events on alluvial fans are infrequent and rarely observed directly. On alluvial fans in some mountainous drainage basins, some flows are often an order of magnitude higher than engineering estimates, even when based on actual measurements of peak sediment and water discharge (*e.g.* Mears, 1979). Information about the nature of these events is usually inferred from the study of sedimentary successions. The identification of the dominant flow processes is especially important for the design of mitigative structures to channel, re-route or dam potentially damaging flows and for the delineation of hazard zones for engineering and planning purposes (Desloges and Gardner, 1984; Jackson, 1987; Costa, 1988). The effectiveness of these remedial measures is dependent on accurate interpretation of flow processes since different flow types show significant differences in rheologic characteristics such as sediment concentration, shear strength, viscosity and turbulence. Thus it is important that sedimentologic criteria for differentiating different types of flows be established and regional variations in dominant flow processes, due to bedrock, climatic or other controls, be identified.

In contrast to fan deposits in many other areas, the Jasper paleofan sediments are characterized by a high proportion of gravel dominated lithofacies. These include very poorly sorted, gravelly diamicton, commonly with numerous boulders (lithofacies Dc, Dci, Dcs, Dcn), and massive to crudely bedded, large pebble to cobble gravel (lithofacies Gm and Gm[a]). The variable sedimentologic characteristics of these gravelly deposits indicates deposition under a wide range of rheologic conditions. Flow types include mainly noncohesive debris flow, hyperconcentrated flood flow and shallow channel, normal stream flow. The abundance of flows dominated by inertial rather than viscous forces, suggests that flow velocities were relatively high (Pierson and Costa, 1987). Flow conditions change both temporally and spatially within and between flow events.

Sediments with characteristics indicative of sieve deposition or cohesive debris flows (mud flows) are rare. Sand and mud lithofacies are also relatively uncommon. The fan successions differ from generalized facies models of alluvial fans in that consistent lateral and vertical facies trends are lacking. In particular, the relative thickness and frequency of debris flow deposits relative to fluvial sediments does not noticeably decrease either down fan (towards the valley centre) or down section at most sites.

The paleofan sediments are inferred to be Mid-Wisconsinan in age as they locally are interbedded with overlying glaciofluvial gravels deposited approximately 29 ka, prior to the advance of Late Wisconsinan ice in the region. Radiocarbon dates of > 39.7 ka and > 48 ka on wood from the upper part of the fan deposits indicates that sedimentation began before that time but the age of the oldest part of the succession is unknown. A Mid-Wisconsinan age for the paleofans may partially account for their more widespread extent, in comparison with modern fans. A cooler climate and less well developed forest cover, compared to the present, were generally characteristic of most Cordilleran regions in Canada during the Mid-Wisconsinan and climatic deterioration was especially pronounced after about 25 ka, reflecting the onset of the last glaciation (Clague and MacDonald, 1989). These conditions probably enhanced slope instabilities and debris flow events that contributed to the extensive development of the paleofans.

The sedimentary characteristics of the paleofan successions are similar to modern fan deposits in the region suggesting deposition under comparable, cool temperate, environmental conditions. Like the paleofans, sedimentation on the modern fans is characterized by episodic, coarse gravelly, noncohesive debris flow deposition. Subsequent winnowing and minor deposition of sands and fine gravels results from intermittent stream activity. Paleodrainage basins probably were small, steep and mostly above tree-line, similar to the modern drainage areas. Most debris flow deposits in the Jasper fans are clay poor, unlike those typical of arid region alluvial fans derived from soils and bedrock rich in clays. Differences in sedimentation probably are also a reflection of more precipitation in the Jasper area than in arid regions. However, well developed stratification and channel-fill deposits, characteristic of glaciofluvial outwash fan gravels (*e.g.* Boothroyd and Ashley, 1975), are lacking. Snow melt is considered an important contributor to spring and summer flood discharges in the Jasper region and may also be a factor resulting in the dominance of non-cohesive debris flows and hyperconcentrated flood flows over cohesive debris flows.

Alluvial fan successions in the Front and Main Ranges are characterized by distinctive facies assemblages. Thickly bedded, cobble to boulder, noncohesive, debris flow deposits are the most prevalent sedimentary facies in the Main Range fans. In contrast, beds or lenses of clast-supported, matrix-filled, bar gravels are commonly associated with pebble to cobble dominated, noncohesive debris flow deposits in the Front Ranges. Interbedded sandy facies and thin trough-shaped lenses of coarse openwork gravel, interpreted as channel deposits, are also common in the Front Ranges. In the Main Ranges, fluvial sands and gravels are relatively rare except at one site where a longitudinally continuous unit of interbedded silts, sands and fine gravels occurs. The latter is interpreted as an usually well preserved back levee deposit formed as a result of deflection of the main feeder channel by a bedrock high. A contrast in clast size, bed thickness and number of beds, provides evidence to suggest that debris flow events were less frequent and of higher magnitude in the Main Ranges than in the Front Ranges. Differences in the sedimentary characteristics of the deposits between the Main and Front Range fans also indicates that, in addition to climatic controls, the bedrock lithology and structure of the source areas has a strong influence on depositional processes. Bedrock geology directly controls the amount and modal size of available detrital sediments and, by controlling basin shape, indirectly effects run-off characteristics. The prevalence of large, although relatively infrequent, bouldery debris flow events in the Main Ranges, relative to the Front Ranges, has important implications for the design of engineering structures and the identification of potential hazard zones on alluvial fans in the Rocky Mountains.

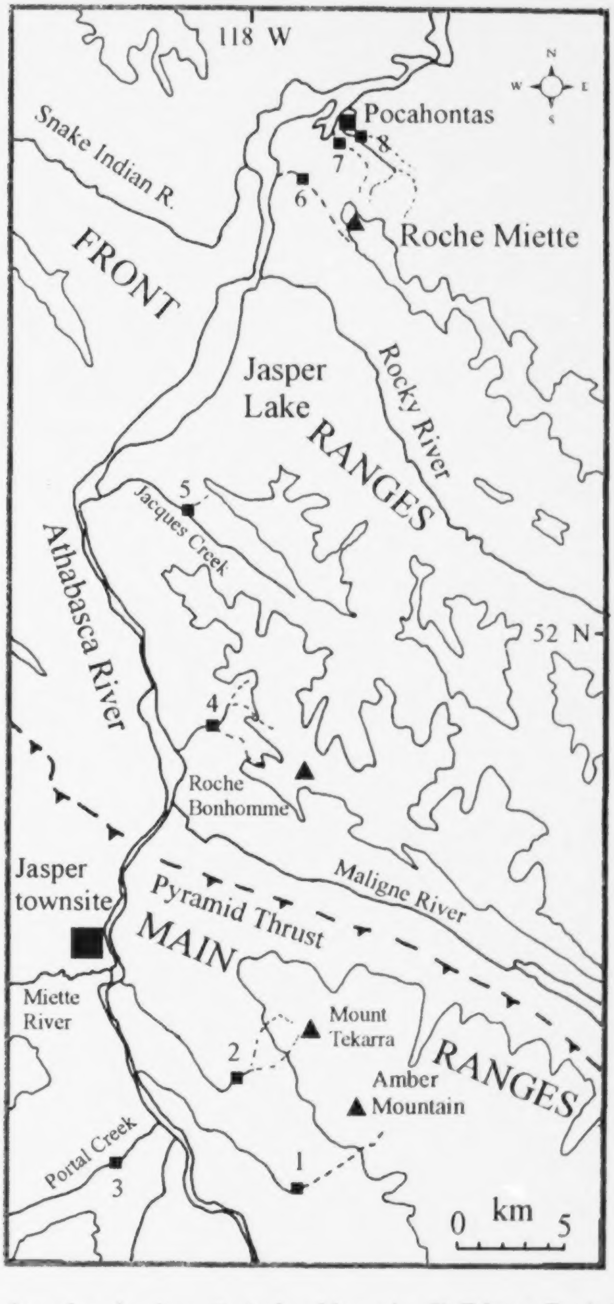

118 W
N
W
E
S
Pocahontas
8
7
6
Snake Indian R.
Roche Miette
FRONT
Jasper
Lake
Rocky River
RANGES
5
Jacques Creek
Athabasca River
52 N
4
Roche
Bonhomme
Jasper
townsite
Pyramid Thrust
Maligne River
MAIN
Miette
River
Mount
Tekarra
2
Amber
Mountain
RANGES
Portal Creek
3
1
0 km 5

Figure 3-1. Location of study sites: 1) Amber Mountain, 2) Tekarra Creek 3) Portal Creek, 4) Roche Bonhomme, 5) Jacques Creek, 6) Roche Miette, 7) Pocahontas, 8) Punchbowl Falls. The 2000 m contour is shown as thin solid lines; dashed lines are intermittent alluvial fan feeder-streams.

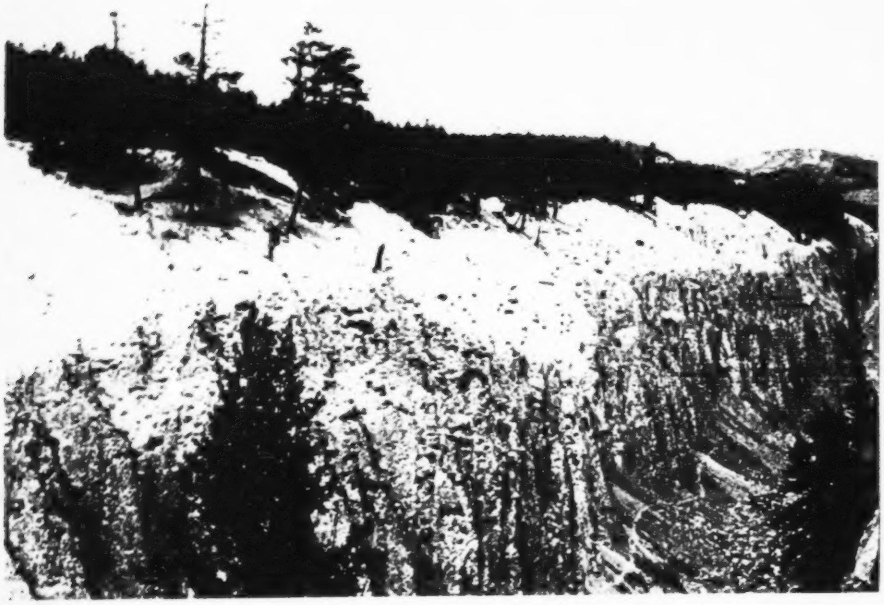


Figure 3-2. a) Bouldery, clast-supported diamicton in the Amber Mountain paleoalluvial fan, unconformably overlain by matrix-supported till (view looking upfan to the northeast); rounded mountain tops at upper right are in the modern fan drainage basin.

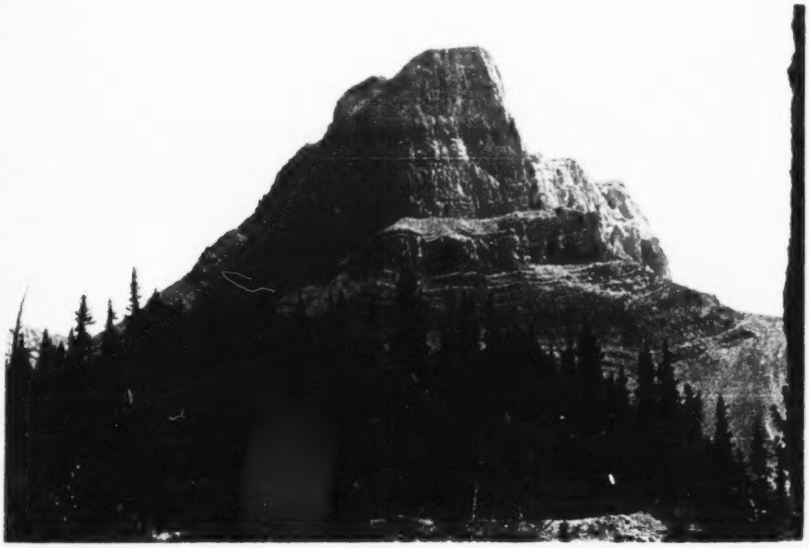


Figure 3-2. b) vertical limestone cliffs and recessive shale units on Roche Miette; tree covered area in foreground is part of the Holocene alluvial fan; the vertical section just visible at the far right of the photo is the base of the Roche Miette paleofan succession.

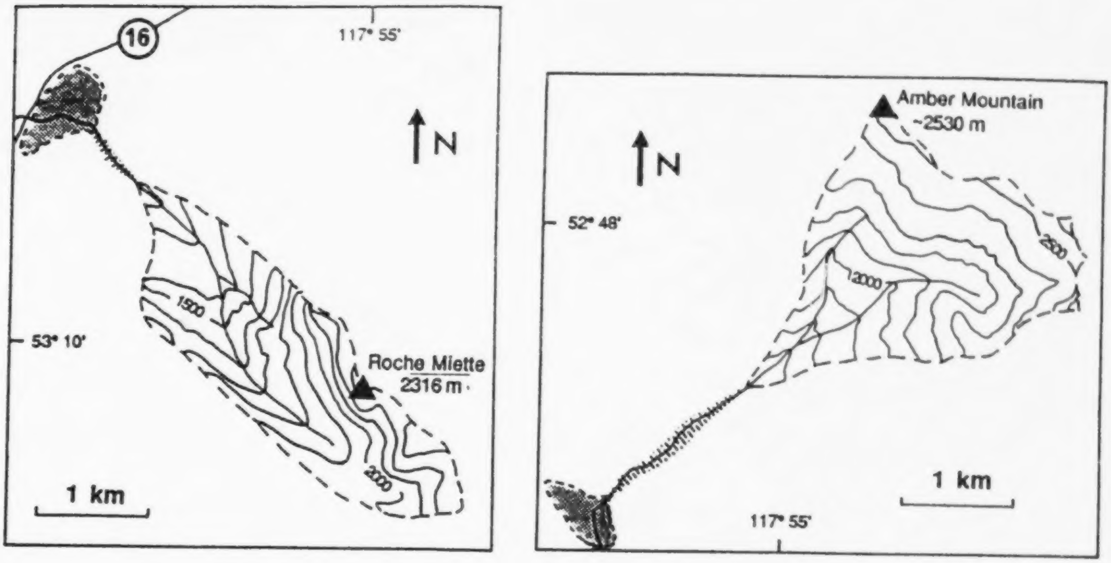

16
117° 55'
N
1500
53° 10'
Roche Miette
2316 m
2000
1 km
Amber Mountain
~2530 m
N
52° 48'
2500
2000
1 km
117° 55'

Figure 3-3. Drainage basin topography of detailed study area: a) Roche Miette fan b) Amber Mountain fan. Shaded areas show approximate extent of Holocene fan sedimentation. Cross-hatched areas show extent of paleoalluvial fan deposits exposed under till in the fan feeder channels.

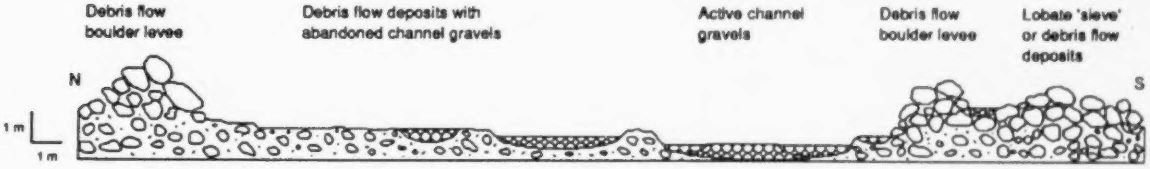

Debris flow
boulder levee
Debris flow deposits with
abandoned channel gravels
Active channel
gravels
Debris flow
boulder levee
Lobate 'sieve'
or debris flow
deposits
N
S
1 m
1 m

Figure 3-4. Diagrammatic cross-section of intermittently active channel on the modern Amber Mountain alluvial fan.

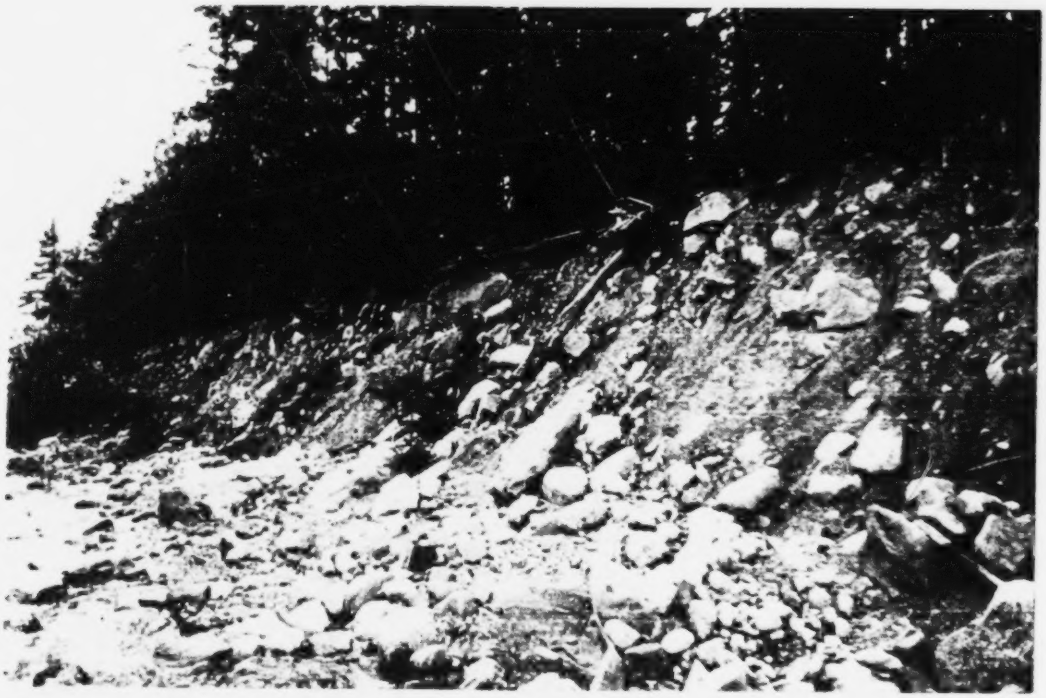


Figure 3-5. Characteristics of recent surficial deposits on the Amber Mountain fan: a) massive to crudely bedded, poorly sorted, gravelly diamicton exposed in a channel wall directly below the fan apex.

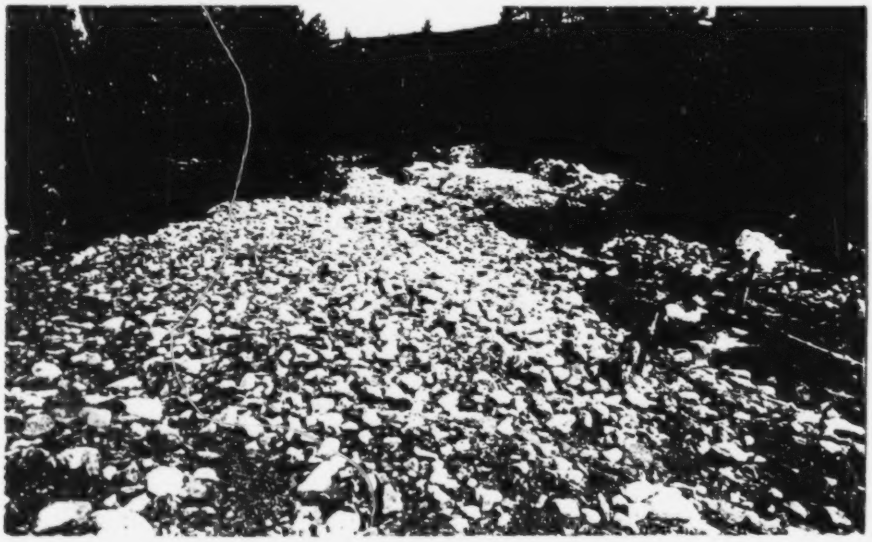


Figure 3-5. b) massive, clast-supported, large pebble to cobble gravel and gravelly diamicton debris flow deposits in the mid-fan region (note handrail of log foot bridge above active channel on right; lower half metre of bridge, exposed at the first site visit, was buried but not eroded by a subsequent debris flow; note also debris flow levees in distance).

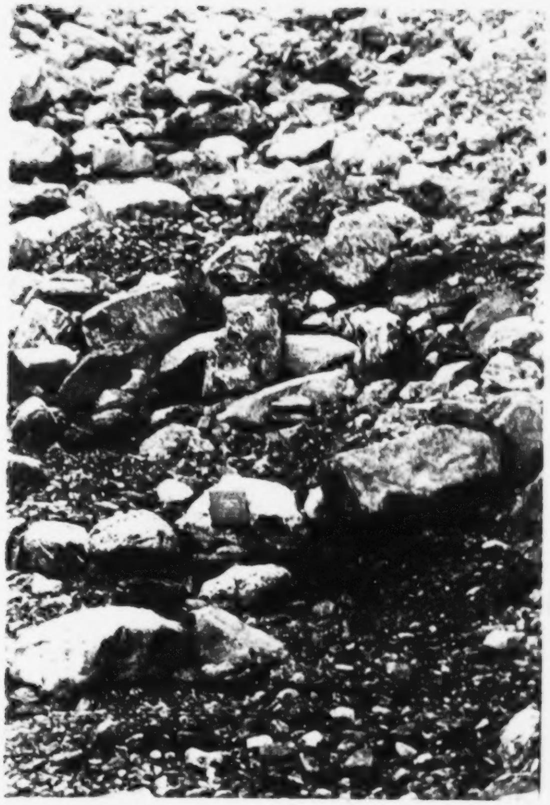


Figure 3-5c) open work, crudely imbricated, pebble and cobble gravels, along base of inactive braid channel.

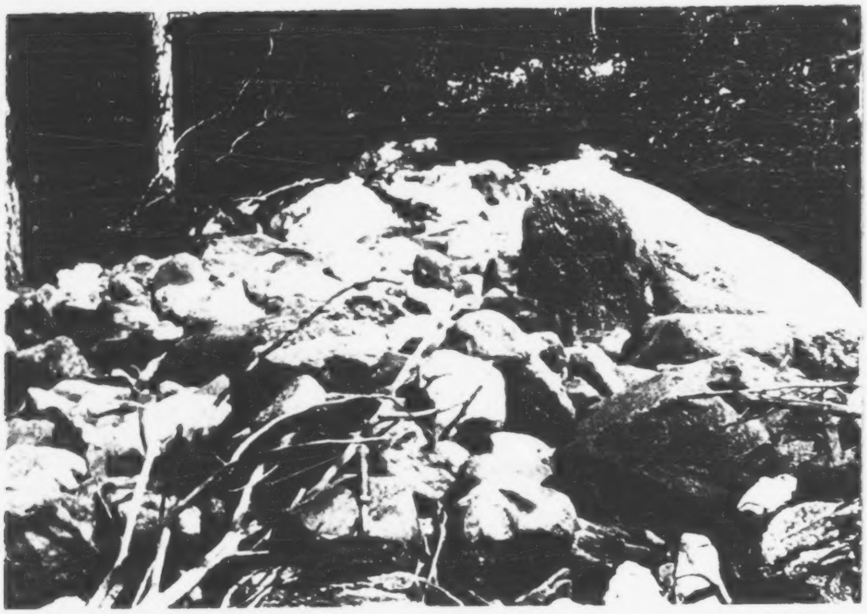


Figure 3-5. d) surface of boulder levee in mid-fan reach (rock hammer in foreground for scale)

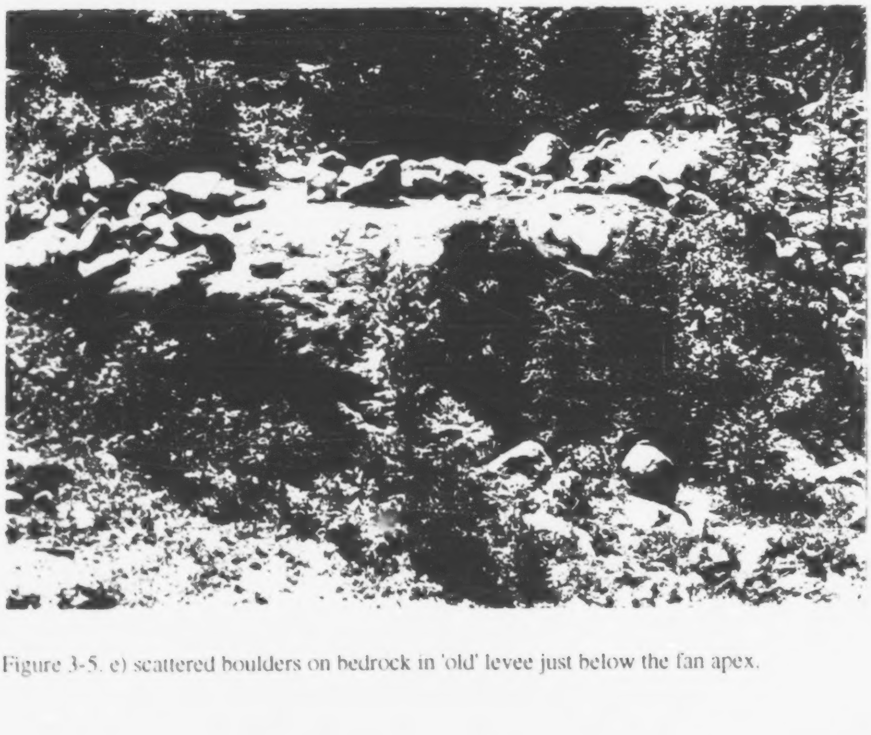


Figure 3-5. e) scattered boulders on bedrock in 'old' levee just below the fan apex.

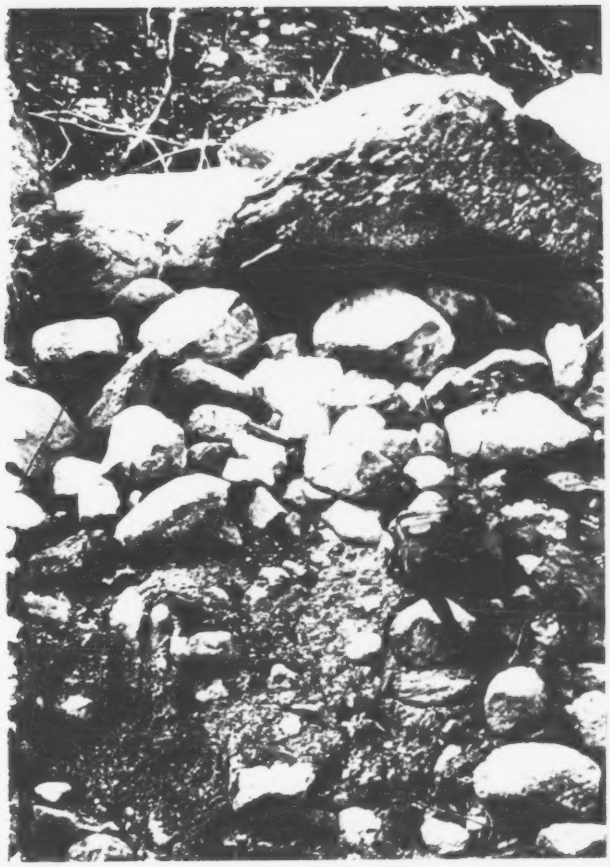


Figure 3-5. f) 'winnowed', open work, gravels overlying poorly sorted, matrix-filled gravelly diamicton exposed in 2 metre high section of boulder levee (camera case is 25 cm high).

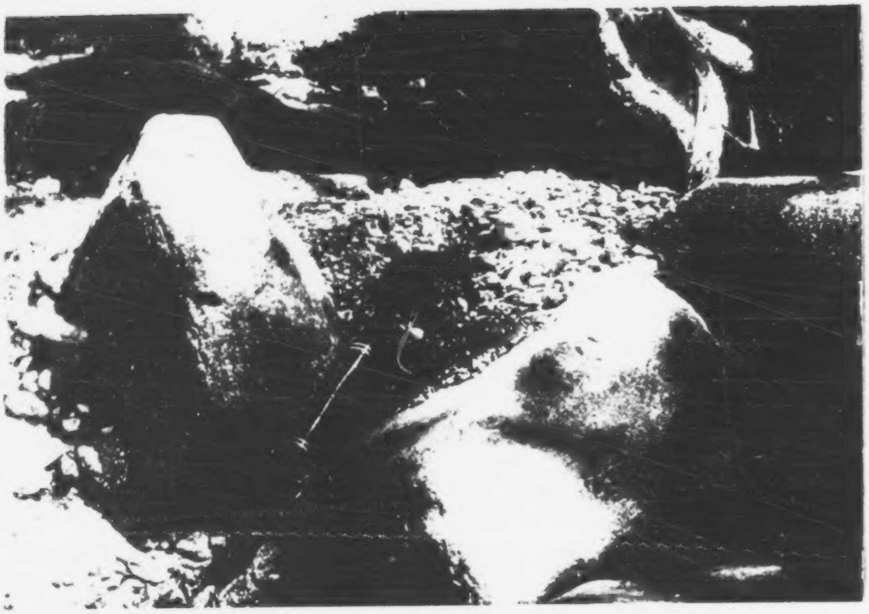


Figure 3-5. g) cobble to boulder gravels with a sorted sandy pebble matrix in the proximal side of a small, lobate deposit that breached the main levee

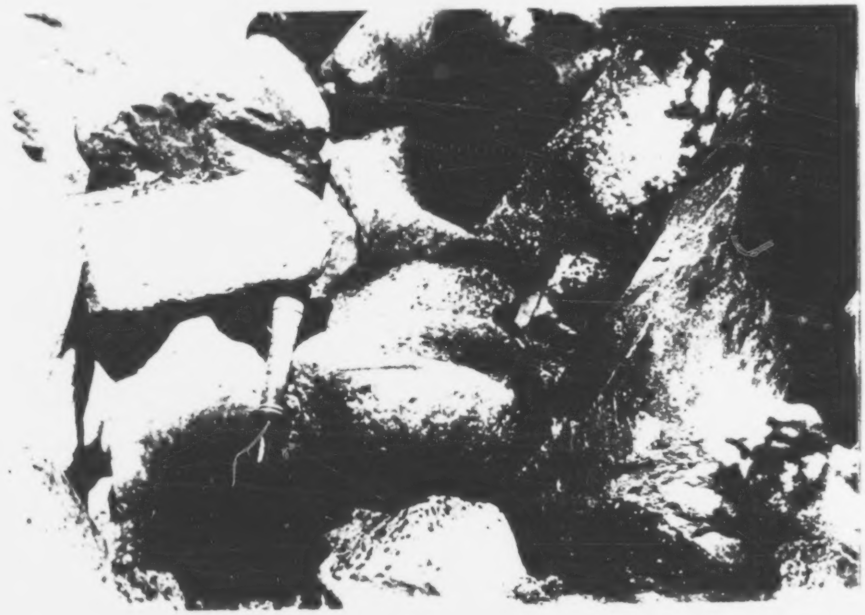


Figure 3-5. h) open work gravels on the distal end of the same gravel lobe shown in g).

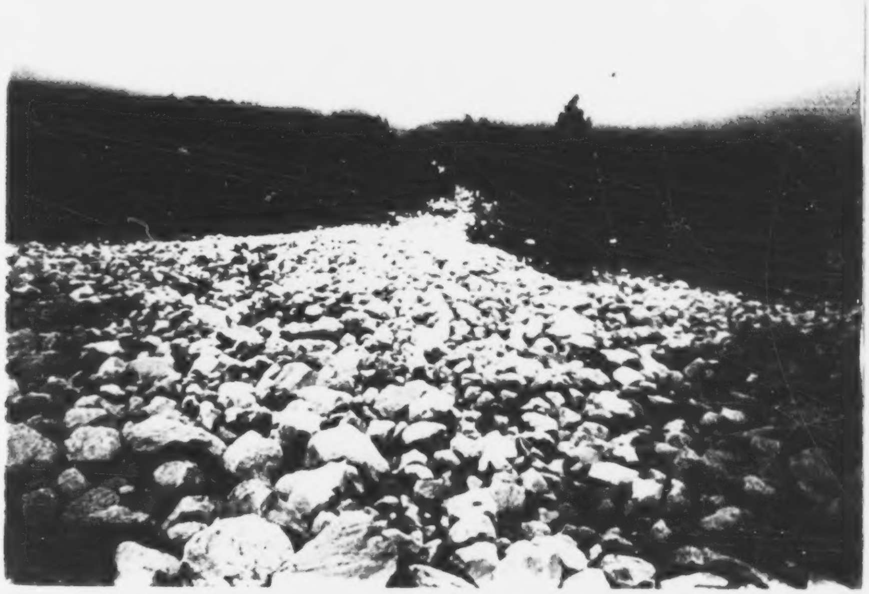


Figure 3-6. Characteristics of recent surficial deposits on the Front Range alluvial fans: a) lobate debris flow deposit that temporarily blocked Highway 93 in northern Banff National Park in 1989 (note the cobble and boulder surface lag, matrix content increases with depth).

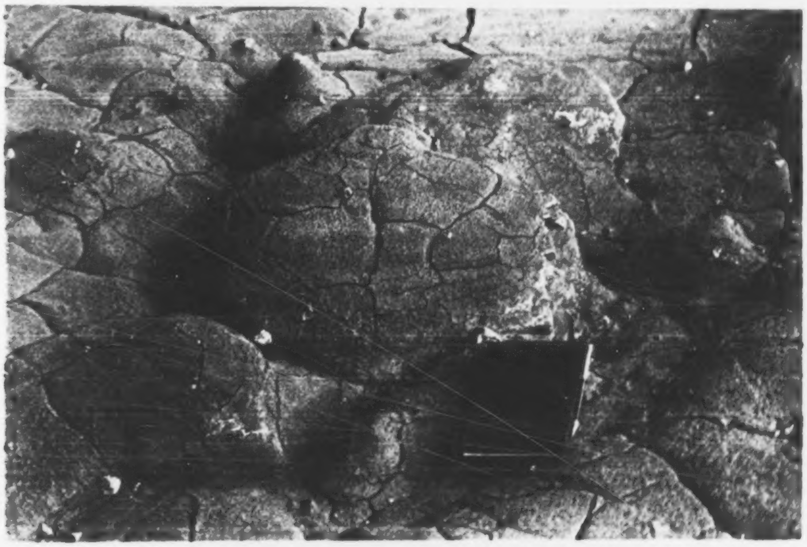


Figure 3-6. b) desiccated mud drapes over boulder gravels in small abandoned channel pool.

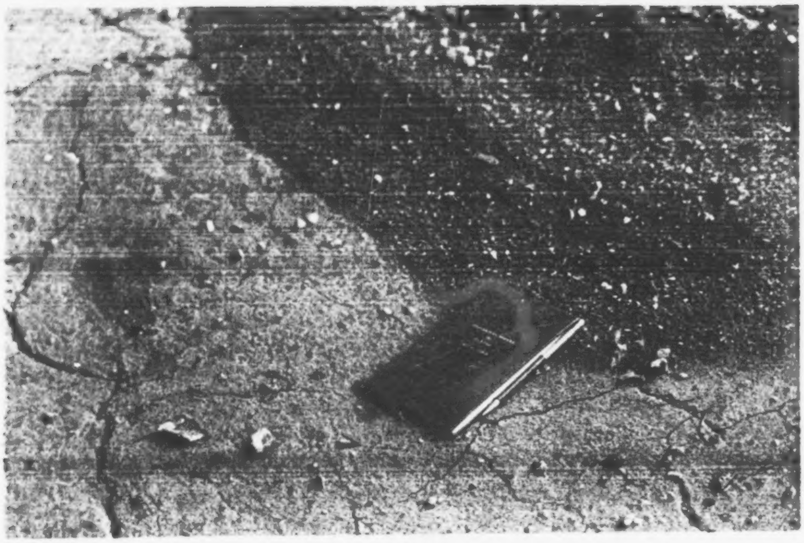


Figure 3-6. c) granular to small pebble gravel deposited over desiccated muds.

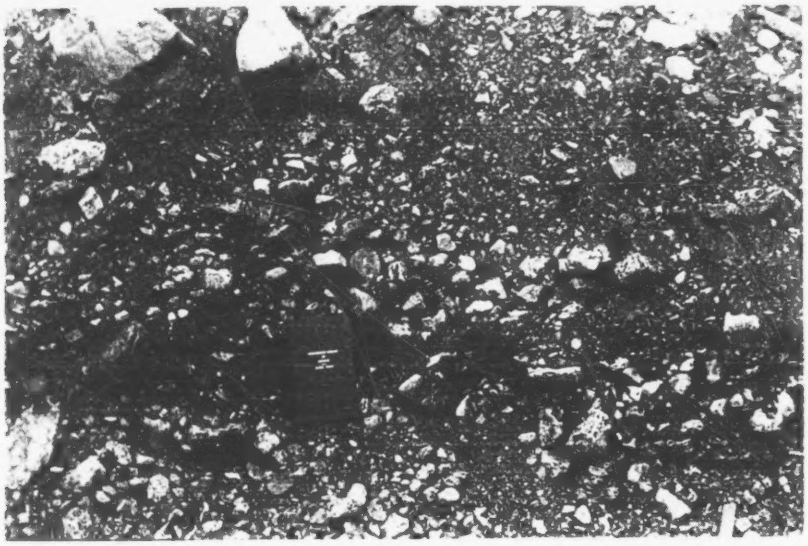


Figure 3-6. d) very crudely bedded, unsorted, muddy to gravelly diamicton exposed in modern fan channel wall (notebook is 21 cm high). Sand and gravel units occur mainly as thin trough-shaped lenses.

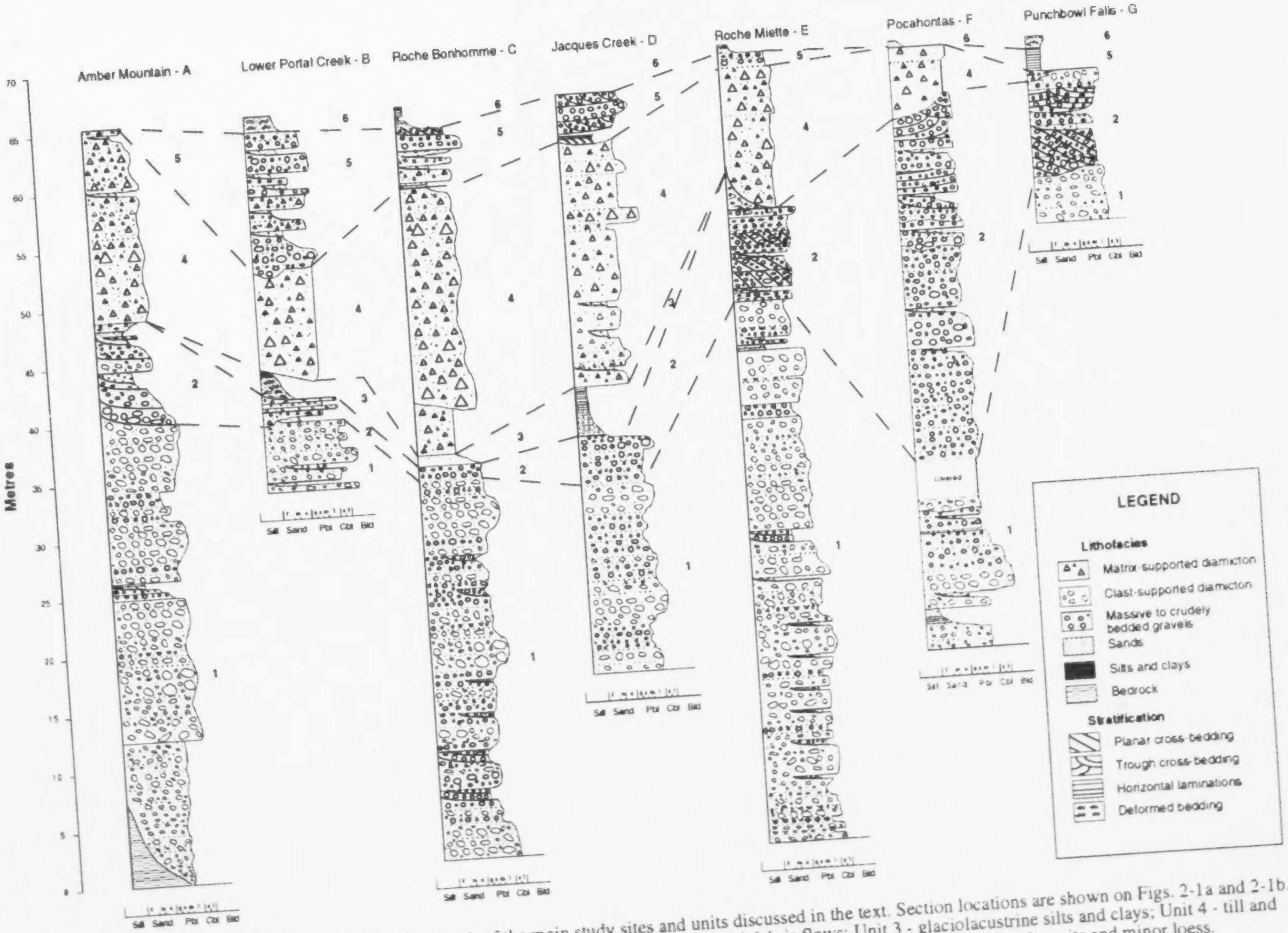

Amber Mountain - A
Lower Portal Creek - B
Roche Bonhomme - C
Jacques Creek - D
Roche Miette - E
Pocahontas - F
Punchbowl Falls - G
Metres
0
5
10
15
20
25
30
35
40
45
50
55
60
65
70
1
2
3
4
5
6
Covered
Silt Sand Pbl Cbl Bld
LEGEND
Lithofacies
Matrix-supported diamicton
Clast-supported diamicton
Massive to crudely bedded gravels
Sands
Silts and clays
Bedrock
Stratification
Planar cross-bedding
Trough cross-bedding
Horizontal laminations
Deformed bedding

Fig. 3-7. Composite stratigraphic columns from each of the main study sites and units discussed in the text. Section locations are shown on Figs. 2-1a and 2-1b. Unit 1 - paleofan deposits; Unit 2 - glaciofluvial gravels and sands and minor proglacial debris flows; Unit 3 - glaciolacustrine silts and clays; Unit 4 - till and associated glacigenic debris flows; Unit 5 - late-glacial glaciolacustrine and glaciofluvial sediments, paraglacial debris flow deposits and minor loess.

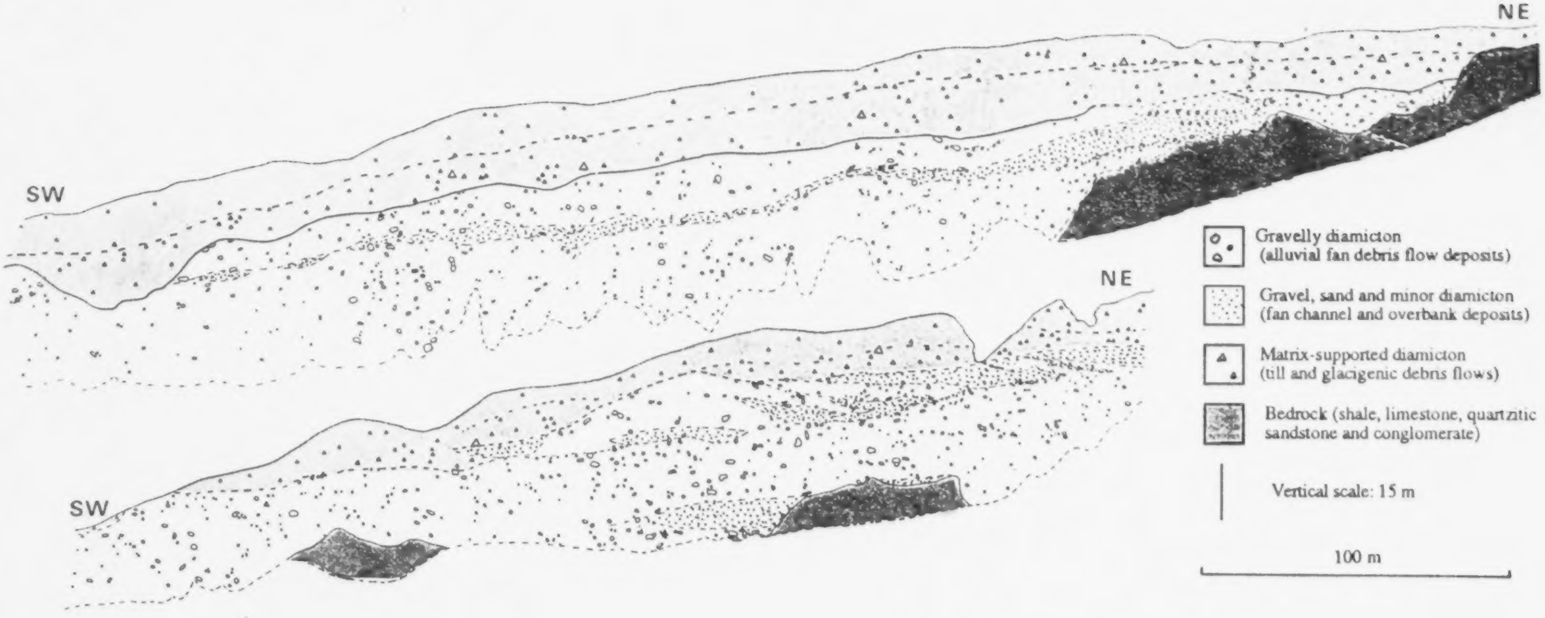

SW
NE
SW
NE
Gravelly diamicton (alluvial fan debris flow deposits)
Gravel, sand and minor diamicton (fan channel and overbank deposits)
Matrix-supported diamicton (till and glacigenic debris flows)
Bedrock (shale, limestone, quartzitic sandstone and conglomerate)
Vertical scale: 15 m
100 m

Figure 3-8. Longitudinal-sections of the Amber Mountain paleofan and associated deposits exposed along the northwest (Figure 3-7a; upper section) and southeast (lower section) walls of the upper fan canyon (reconstructed from moving station panoramic photographs - otherwise uncorrected for parallax error). Symbols are proportional to clast size and represent actual distribution.

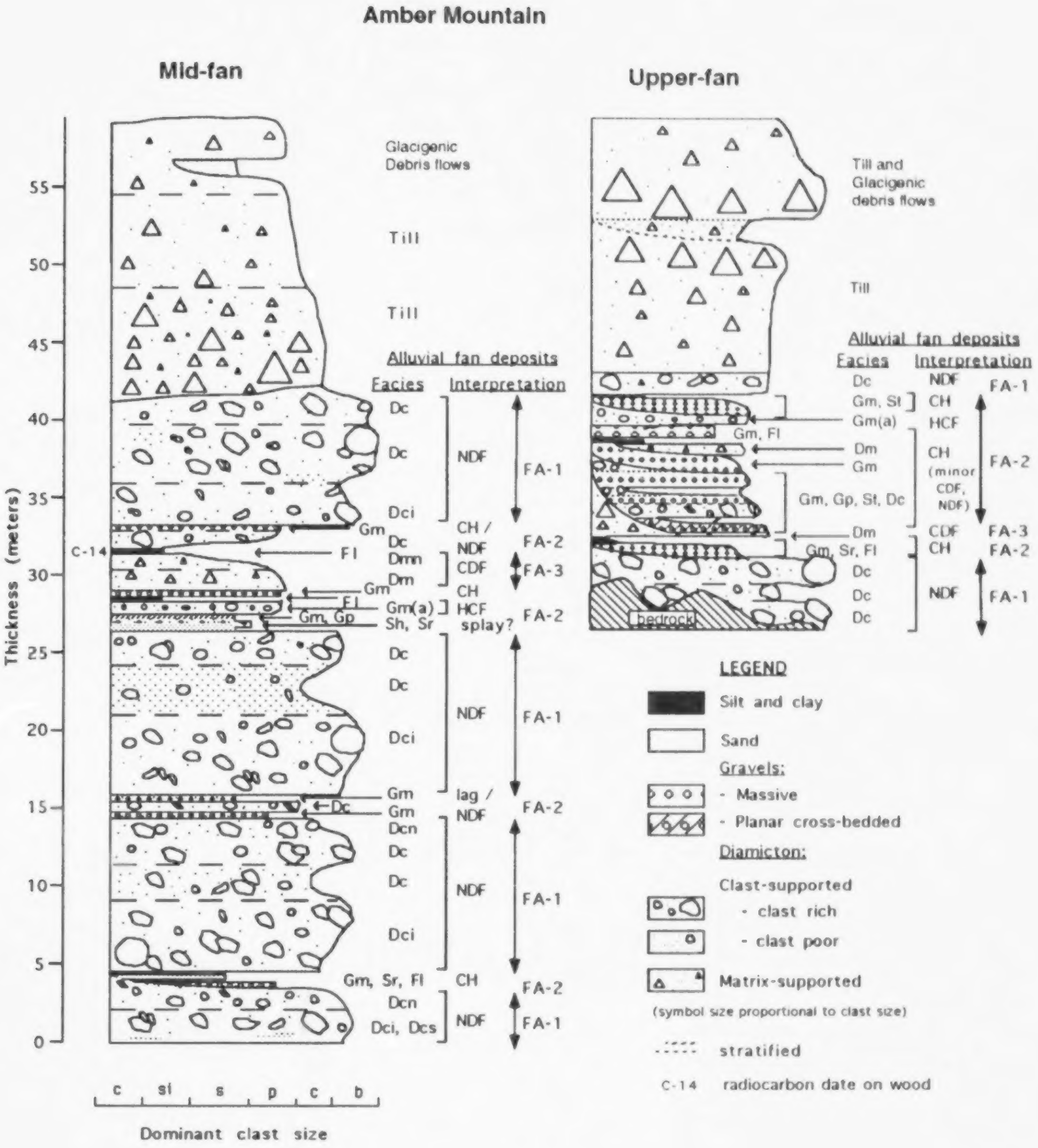

Amber Mountain
Mid-fan
Upper-fan
Glacigenic Debris flows
Till
Till
Alluvial fan deposits
Facies
Interpretation
Dc
Dc
Dci
NDF
FA-1
Gm
Dc
CH /
NDF
FA-2
C-14
Fl
Dmm
Dm
CDF
FA-3
Gm
CH
Fl
Gm(a)
HCF
Gm, Gp
Sh, Sr
splay?
FA-2
Dc
Dc
Dci
NDF
FA-1
Gm
lag /
Dc
Gm
NDF
FA-2
Dcn
Dc
Dc
Dci
NDF
FA-1
Gm, Sr, Fl
CH
FA-2
Dcn
Dci, Dcs
NDF
FA-1
Thickness (meters)
0
5
10
15
20
25
30
35
40
45
50
55
c
si
s
p
c
b
Dominant clast size
Till and Glacigenic debris flows
Till
Alluvial fan deposits
Facies
Interpretation
Dc
NDF
FA-1
Gm, St
CH
Gm(a)
HCF
Gm, Fl
Dm
Gm
CH (minor CDF, NDF)
FA-2
Gm, Gp, St, Dc
Dm
CDF
FA-3
Gm, Sr, Fl
CH
FA-2
Dc
Dc
Dc
NDF
FA-1
bedrock
LEGEND
Silt and clay
Sand
Gravels:
- Massive
- Planar cross-bedded
Diamicton:
Clast-supported
- clast rich
- clast poor
Matrix-supported
(symbol size proportional to clast size)
stratified
C-14 radiocarbon date on wood

Figure 3-9. Detailed sedimentologic logs of measured sections at Amber Mountain (Site 1). Abbreviations as follows: NDF - noncohesive debris flow; CDF - cohesive debris flow; HCF - hyperconcentrated flood flow; GF - Grain flow (sieve deposits); CH - channel fill succession; lag - channel lag gravels; bar - longitudinal bar and gravel sheet deposits; FA1 to FA3 - facies associations 1 to 3. See text for explanation of lithofacies.

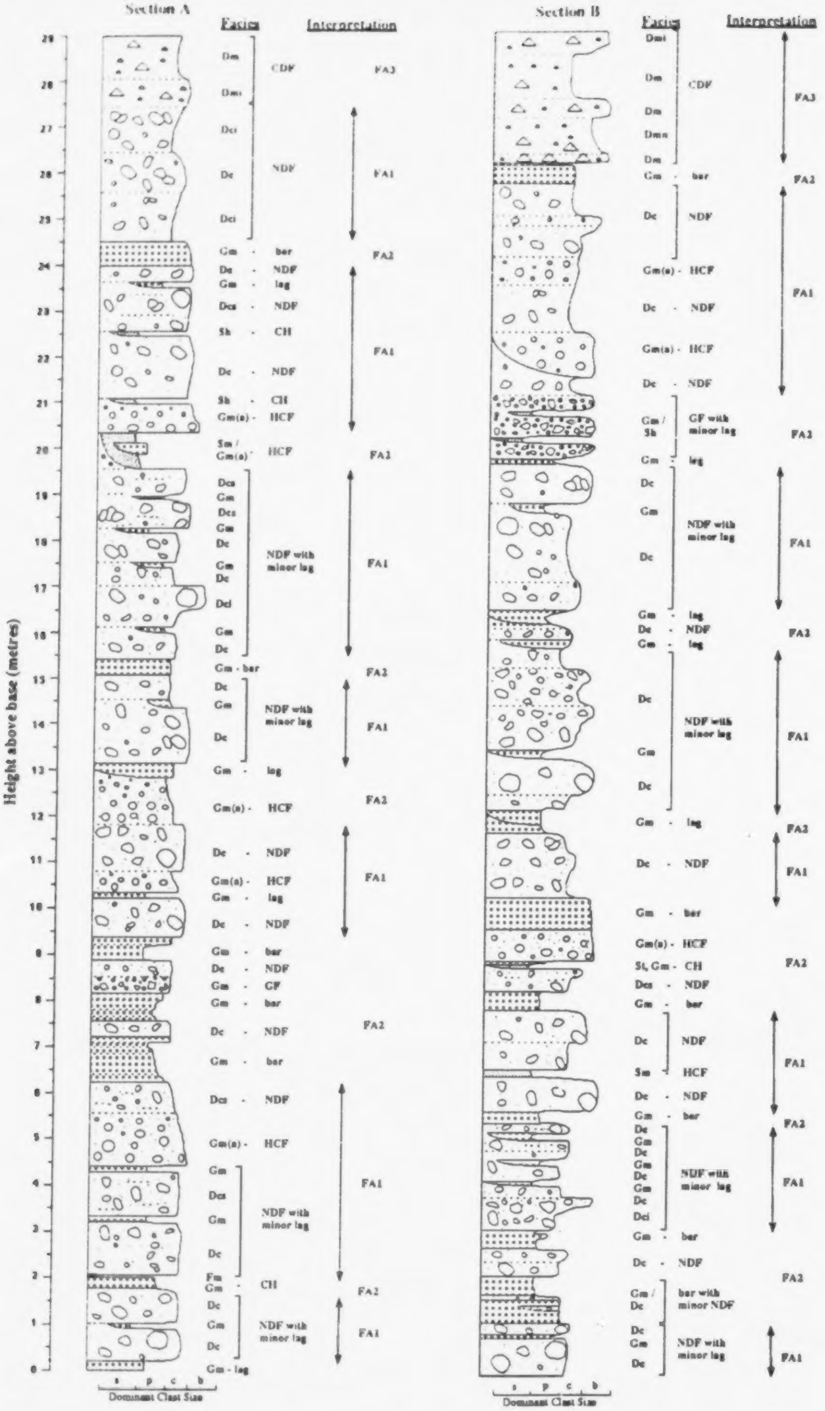

Section A
Facies
Interpretation
Height above base (metres)
29
28
27
26
25
24
23
22
21
20
19
18
17
16
15
14
13
12
11
10
9
8
7
6
5
4
3
2
1
0
Dm
Dmi
Dci
Dc
Dci
CDF
NDF
FA3
FA1
Gm - bar
Dc - NDF
Gm - lag
Dcs - NDF
Sh - CH
Dc - NDF
Sh - CH
Gm(s) - HCF
Sm / Gm(s) HCF
FA2
FA1
FA2
Dcs
Gm
Dcs
Gm
Dc
Gm
Dc
Dcl
Gm
Dc
NDF with minor lag
FA1
Gm - bar
FA2
Dc
Gm
Dc
NDF with minor lag
FA1
Gm - lag
Gm(s) - HCF
FA2
Dc - NDF
Gm(s) - HCF
Gm - lag
Dc - NDF
FA1
Gm - bar
Dc - NDF
Gm - GF
Gm - bar
Dc - NDF
Gm - bar
FA2
Dcs - NDF
Gm(s) - HCF
Gm
Dcs
Gm
Dc
NDF with minor lag
FA1
Fm
Gm - CH
FA2
Dc
Gm
Dc
NDF with minor lag
FA1
Gm - lag
s p c b
Dominant Clast Size
Section B
Facies
Interpretation
Dmi
Dm
Dm
Dmn
Dm
CDF
FA3
Gm - bar
FA2
Dc
NDF
Gm(s) - HCF
Dc - NDF
Gm(s) - HCF
Dc - NDF
FA1
Gm / Sh
GF with minor lag
Gm - lag
FA2
Dc
Gm
Dc
NDF with minor lag
FA1
Gm - lag
Dc - NDF
Gm - lag
FA2
Dc
Gm
Dc
NDF with minor lag
FA1
Gm - lag
FA2
Dc - NDF
FA1
Gm - bar
Gm(s) - HCF
St, Gm - CH
Dcs - NDF
Gm - bar
FA2
Dc
NDF
Sm
HCF
Dc - NDF
FA1
Gm - bar
FA2
Dc
Gm
Dc
Gm
Dc
Gm
Dc
Dci
NDF with minor lag
FA1
Gm - bar
Dc - NDF
Gm / Dc
bar with minor NDF
FA2
Dc
Gm
Dc
NDF with minor lag
FA1
s p c b
Dominant Clast Size

Figure 3-10. Detailed sedimentologic logs of measured sections at Roche Miette. Section A is located 115 metres up-fan of section B. Section C is a detailed log of the lower 5 metres of the section illustrated in Figure 3-7E, located approximately half way between sections A and B. See Figure 3-9 for legend and explanation of abbreviations. (Note the difference in vertical scale when comparing Figures 3-9 and 3-10).

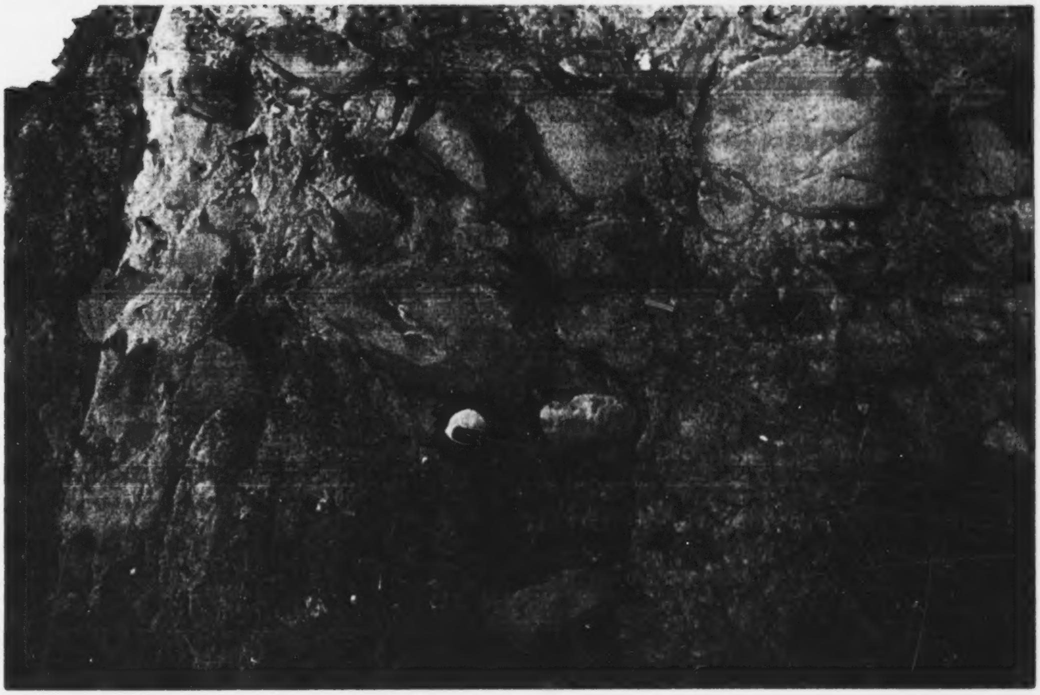


Figure 3-11. Gravelly diamicton (facies Dc, Dcs): a) massive, mainly clast-supported, gravelly diamicton; note poor sorting, crude imbrication of large clasts and clustering of clasts upflow (right) of large boulder.

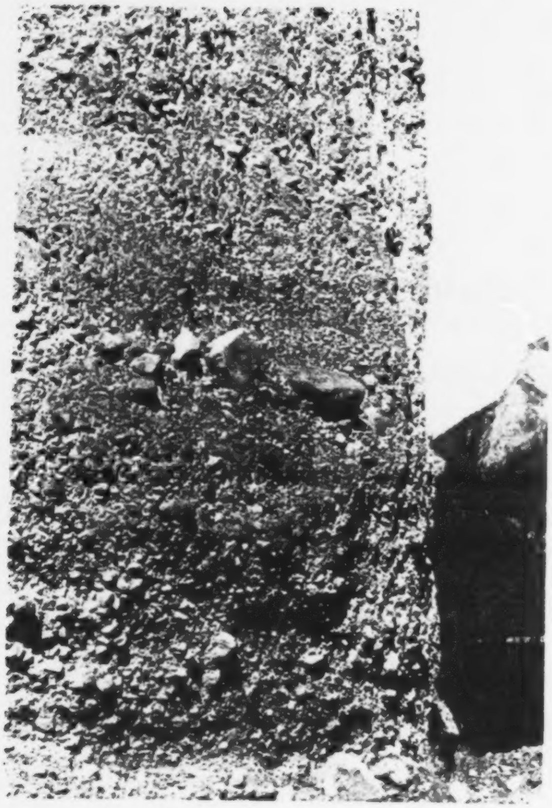


Figure 3-11. b) clast-supported cobble to boulder diamicton (facies Dc) interbedded with massive to crudely bedded pebble gravels (facies Gm); note poorly defined bed contacts (large boulder at photo centre is about 60 centimetres in diameter).

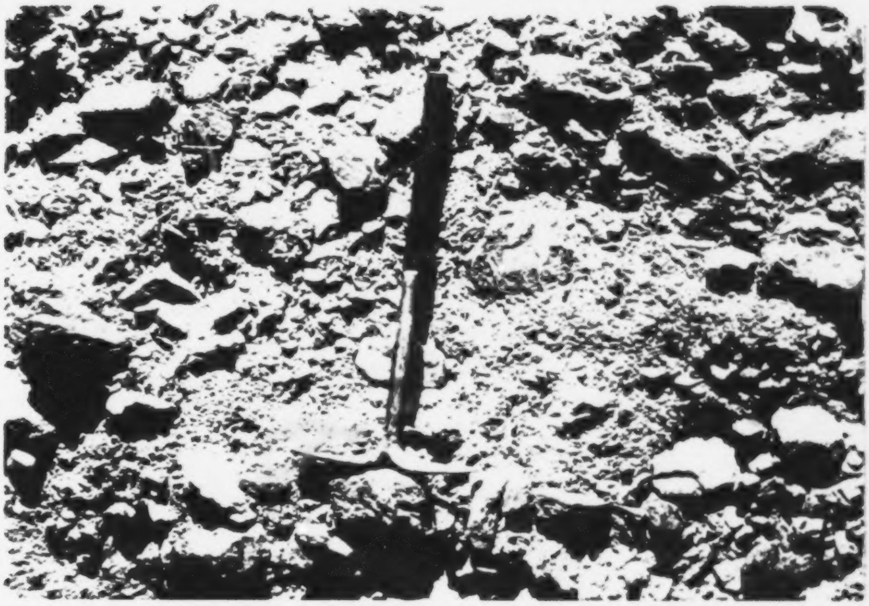


Figure 3-11. c) cobble clasts 'floating' in a disorganized sand-pebble matrix in a 'matrix'-supported gravelly diamicton (pick is 65 centimetres long).

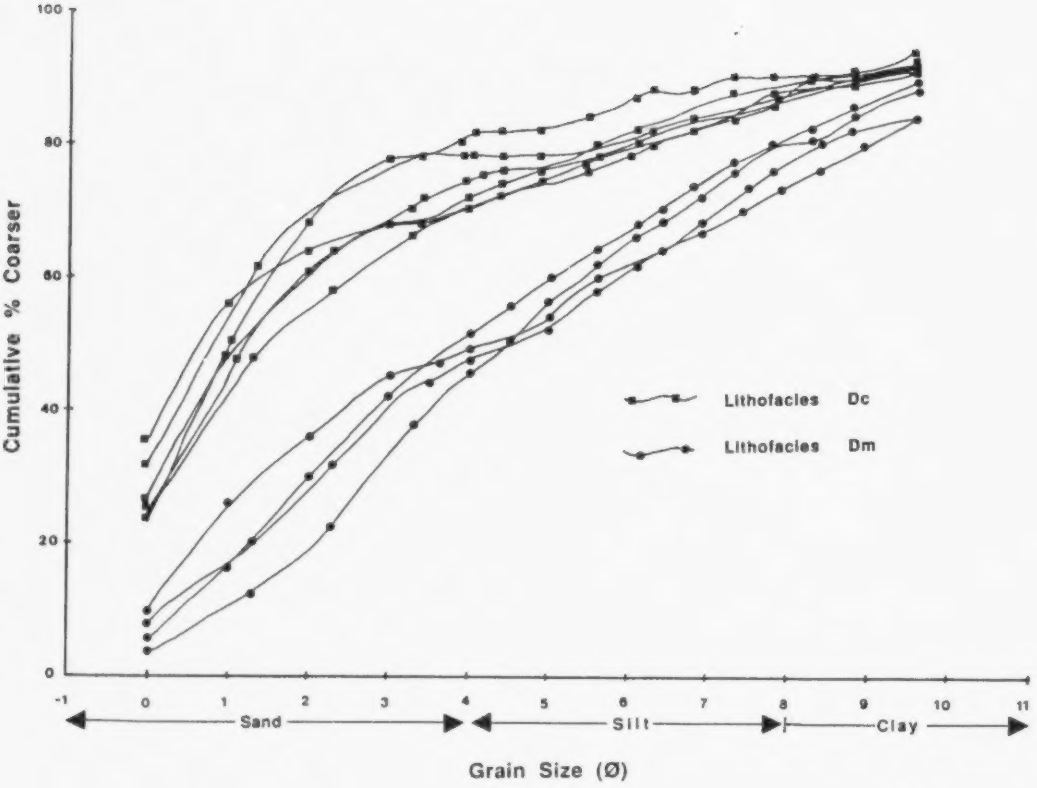


Figure 3-12. Cumulative grain-size distribution curves of matrix textures (finer than 1 mm) from representative samples of lithofacies Dc and Dm.

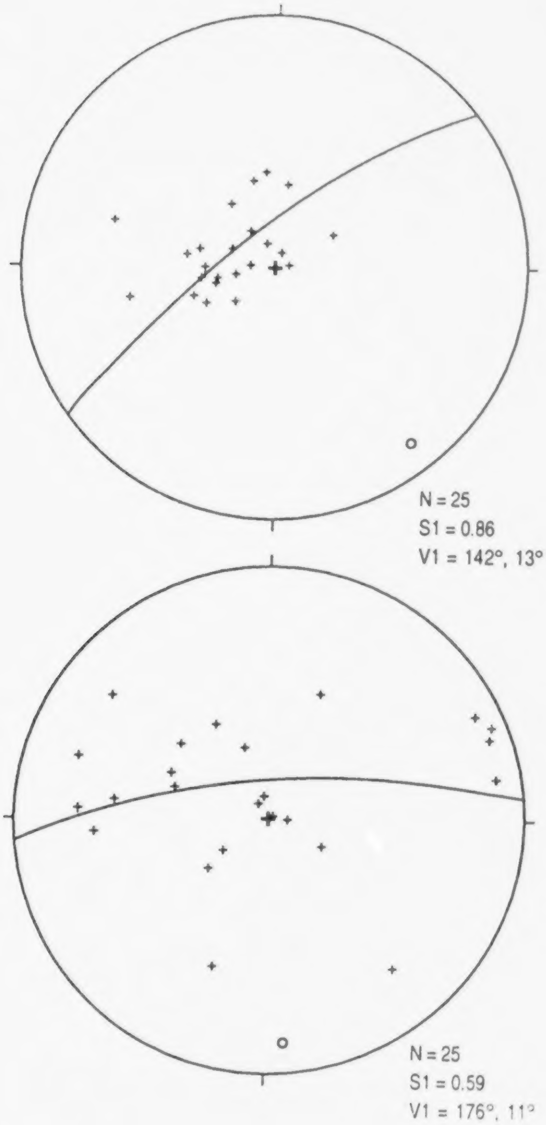


Figure 3-13. Lower hemisphere projections of poles to a-b planes of tabular clasts in massive gravel (facies Gm - top) and gravelly diamicton (facies Dc - bottom). Open dots are poles to best-fit great circles. Paleoflow is to the northwest in both cases.

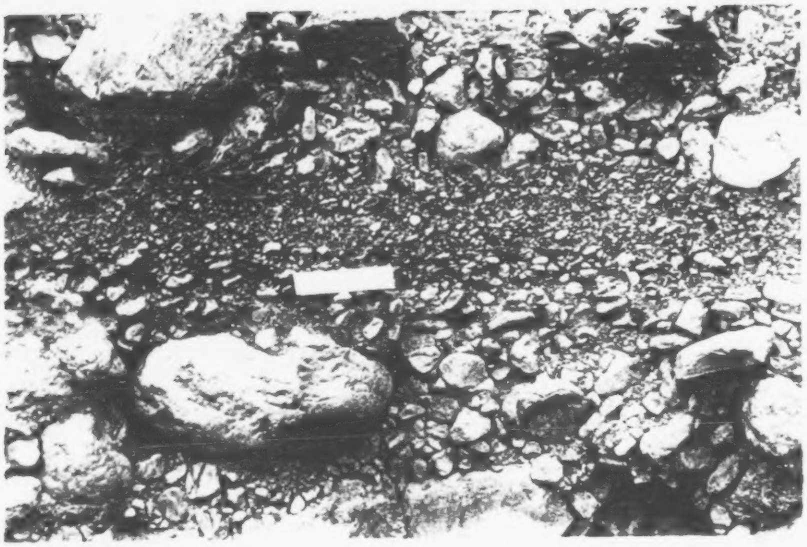


Figure 3-14. Gravelly lithofacies: a) interbedded pebble and cobble gravels (facies Gm), note imbrication - paleoflow to the right.

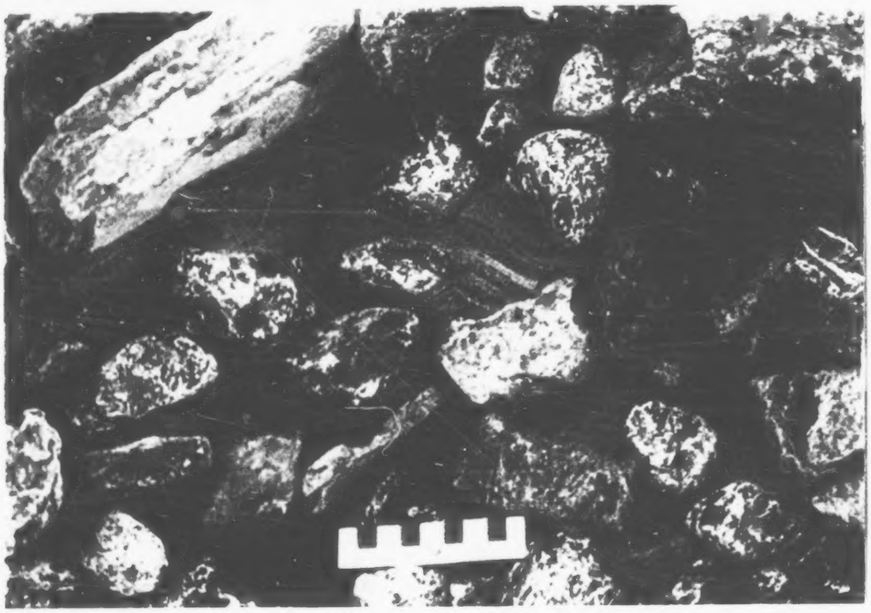


Figure 3-14. b) bimodal, large pebble gravel with stratified sand matrix; note open work gravels at top of bed.

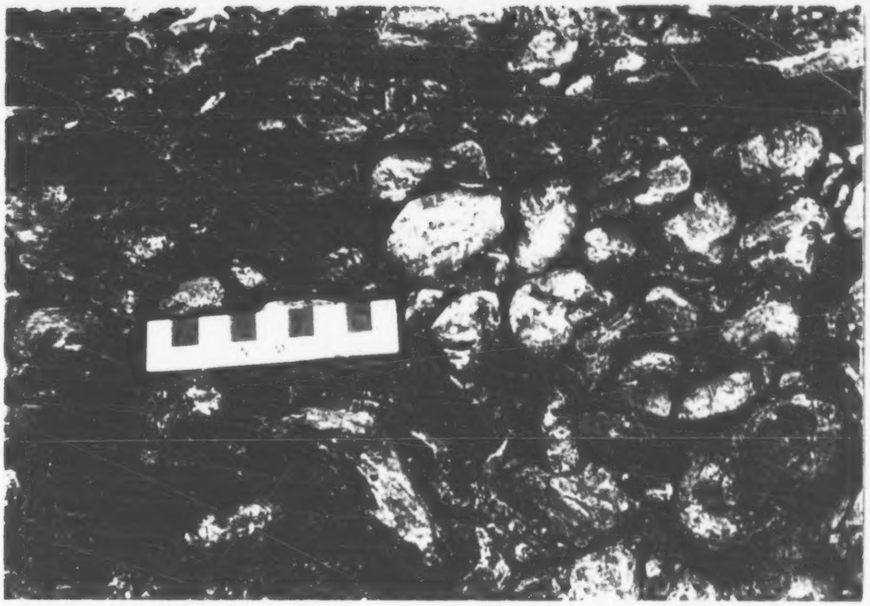
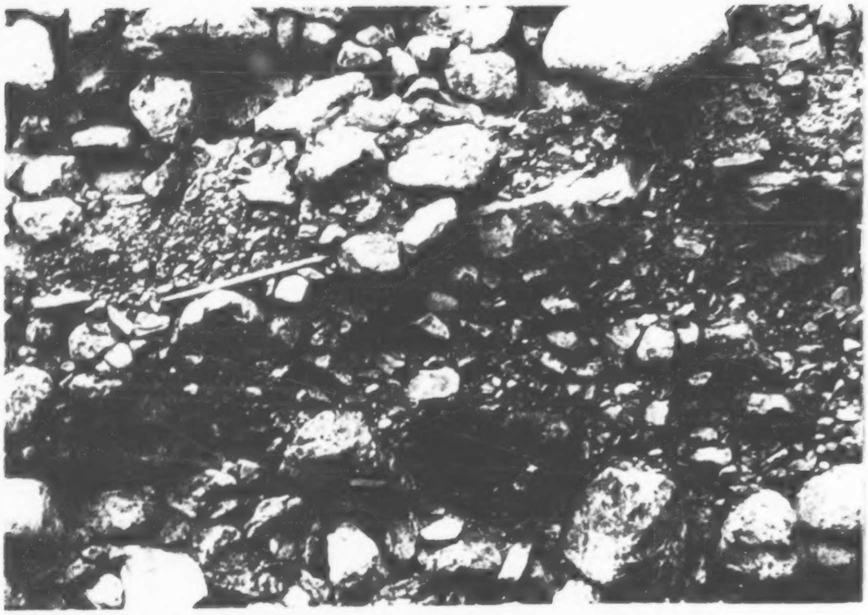


Figure 3-14. c) rounded, imbricated, openwork gravels of facies Gm; dark colour is due to manganese and iron oxide-staining.

Figure 3-14. d) thin trough-shaped lenses, 1-5 metres wide, of openwork gravels (dark coloured lenses) interbedded with massive gravels (Facies Gm); strata dip a few degrees down-valley.



3-14 e) coarse 'lag' gravels at the base of thin trough shaped gravel lenses (pencil marks small pebble lag gravel in the middle of three beds; trowel is 20 centimetres long).

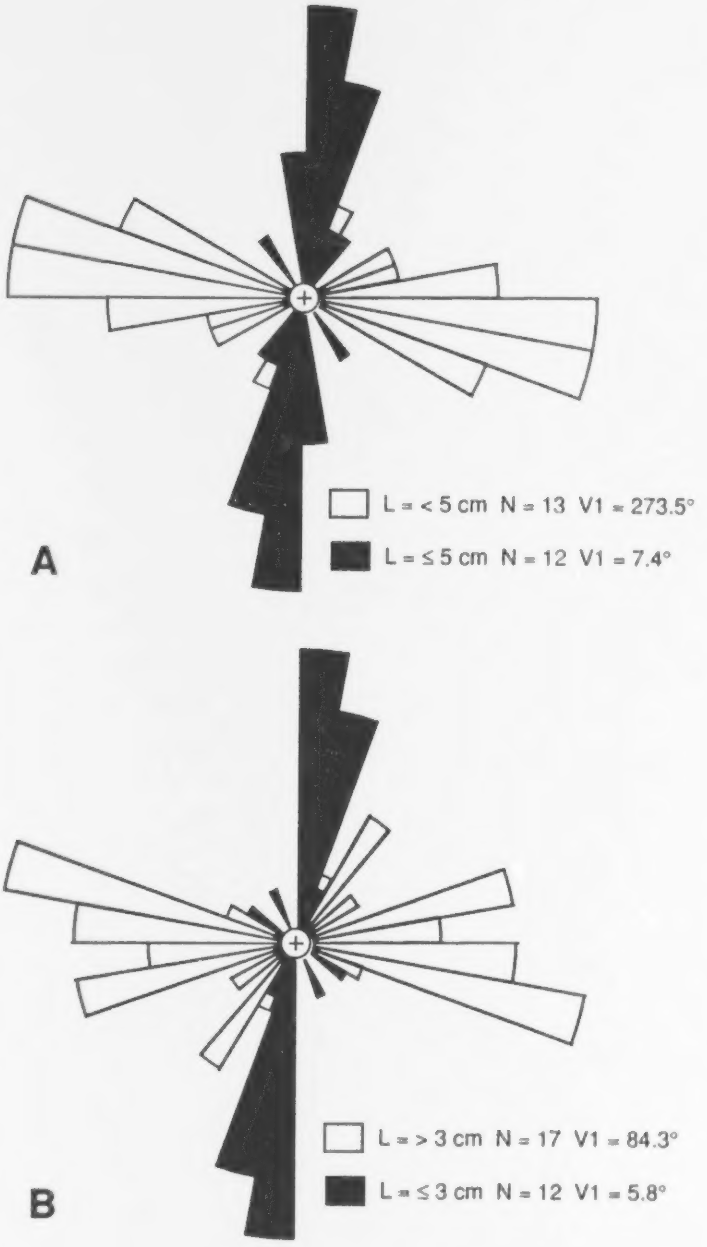


Figure 3-15. Rose diagrams of a-axis orientations in massive gravel beds of facies Gm(a) at A) Pocahontas and B) Roche Miette. Paleoflow direction is northerly at both sites. The bimodal fabric is controlled by clast size with cobbles and boulders showing mainly a-axis transverse-to-flow orientations and pebbles displaying mostly a-axis parallel-to-flow alignments.

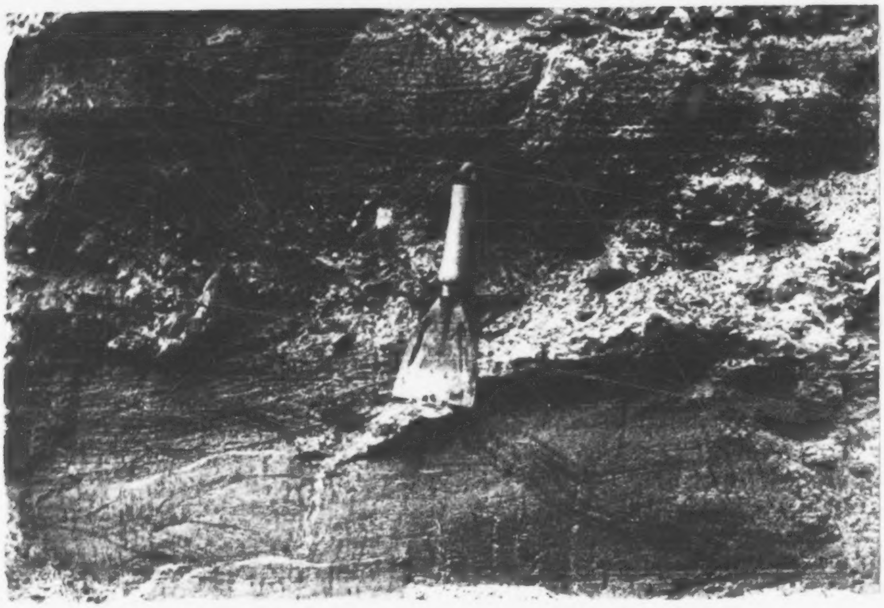


Figure 3-16. Fine-grained lithofacies: a) parallel laminated silt and fine sand (facies Fl) and trough cross-laminated sands (St) with interbedded sandy gravel bed (centre); laminae in lower sand are overprinted by groundwater iron staining (below trowel).

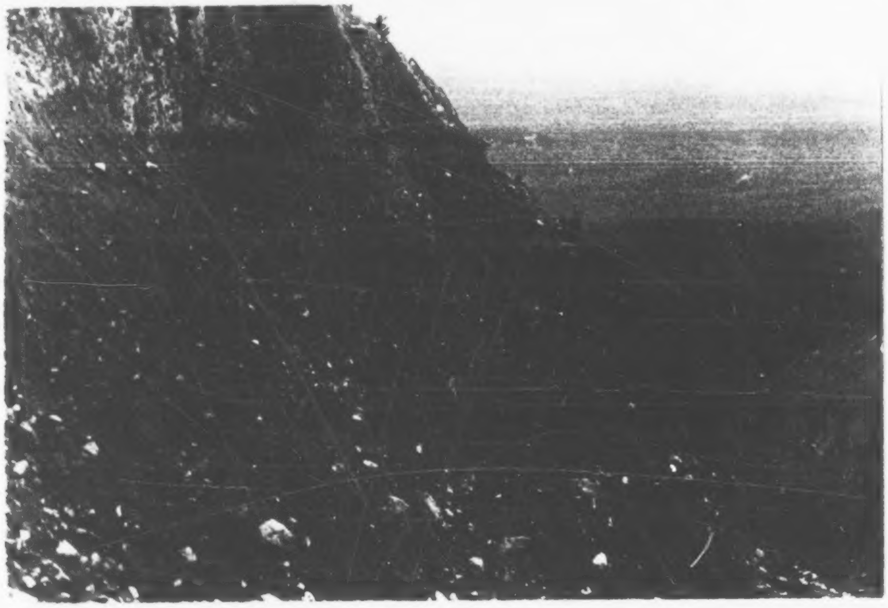


Figure 3-16. b) channel sands and interbedded poorly sorted coarse gravels; note lag gravel (bouldery bed) at base of lower channel.

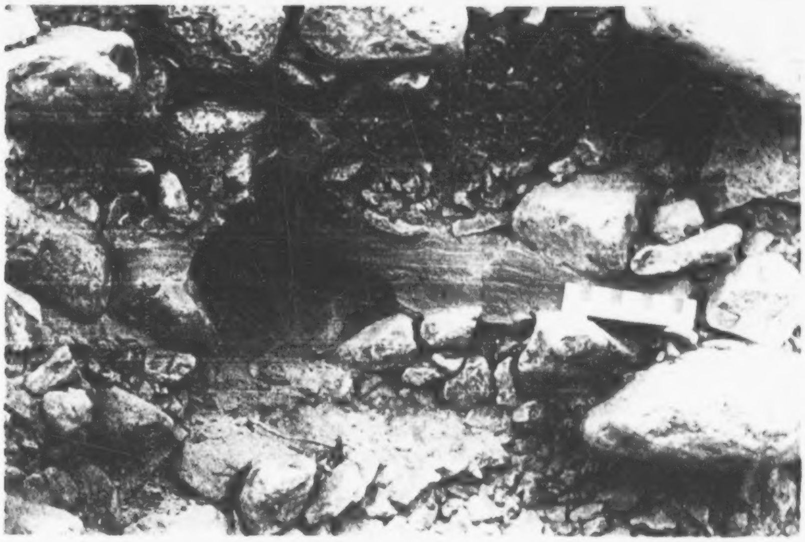


Figure 3-16. c) small sand lense with poorly developed draped laminae (facies Sr) interbedded with openwork gravels; parallel laminated fine sands and silts (facies Fl) infill low areas between clasts at the base of the lense.

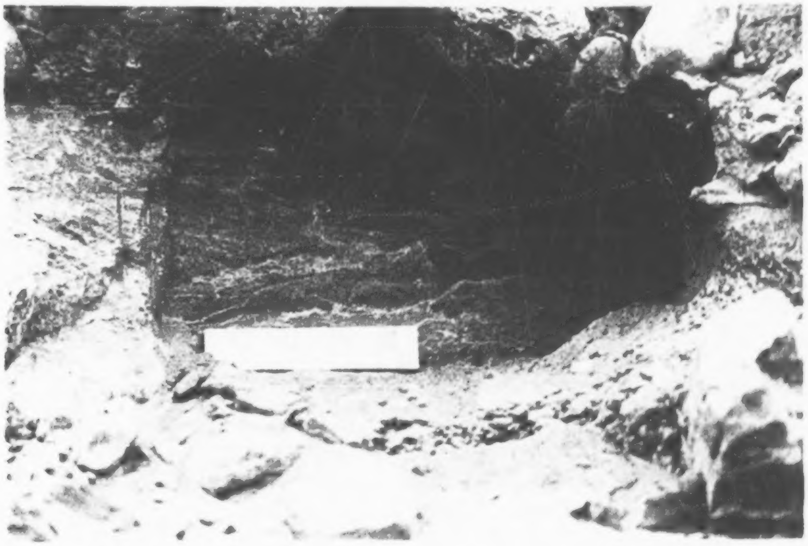


Figure 3-16. d) fluid-escape and loading structures in small sand lense.

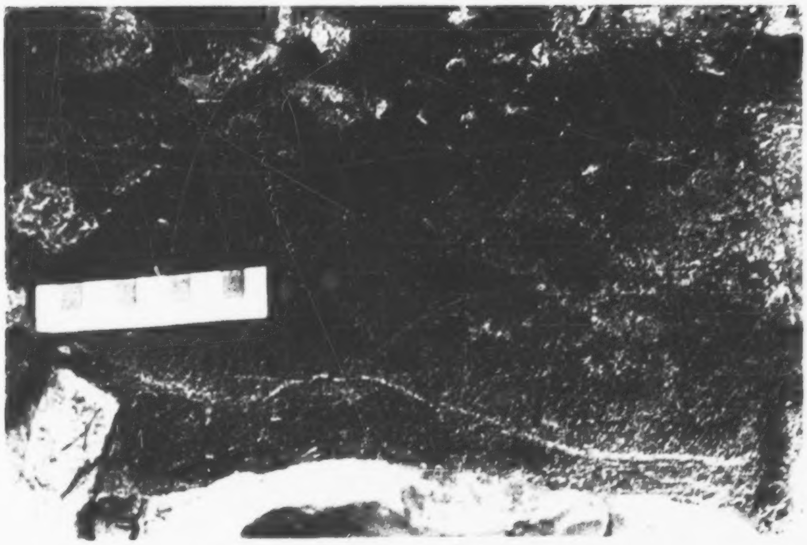


Figure 3-16. e) dish structures and loaded upper contact in sand lense; note preserved fine sand laminae at base of lense.

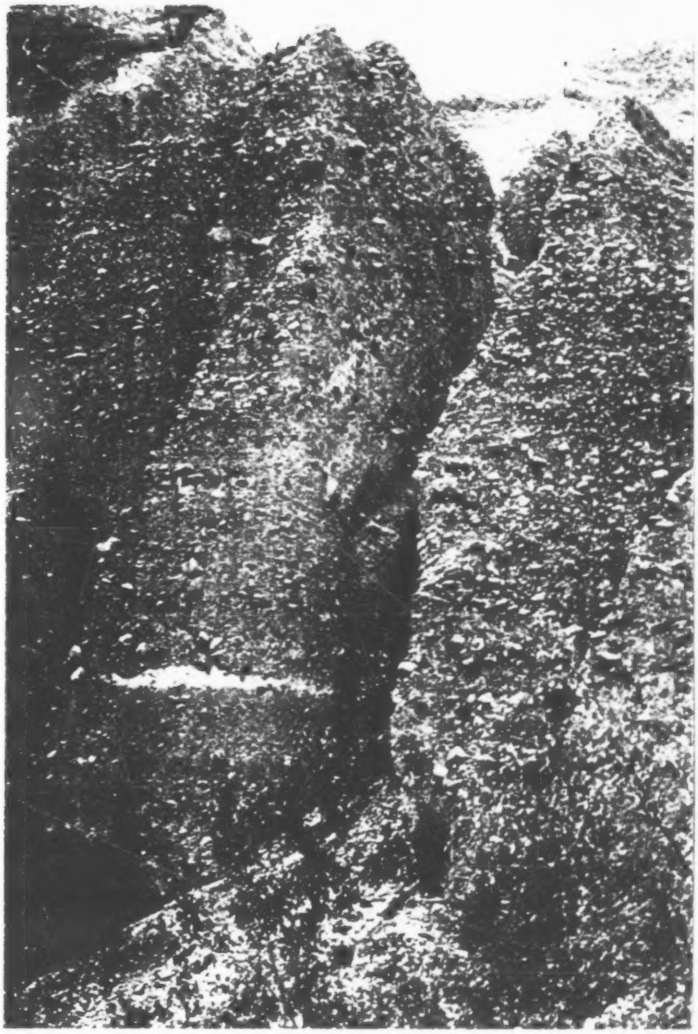


Figure 3-16. f) lense of massive sand (facies Sm) 40 centimetres thick and 7 metres wide (man for scale in lower centre).

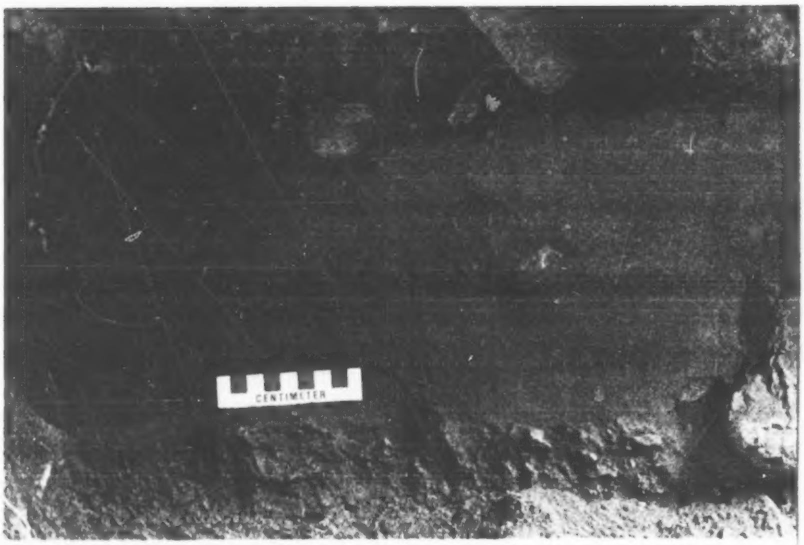


Figure 3-16. g) angular, isolated pebble in thin massive sand bed; note sharp, loaded upper contact with overlying matrix-supported diamicton (facies Dmm).

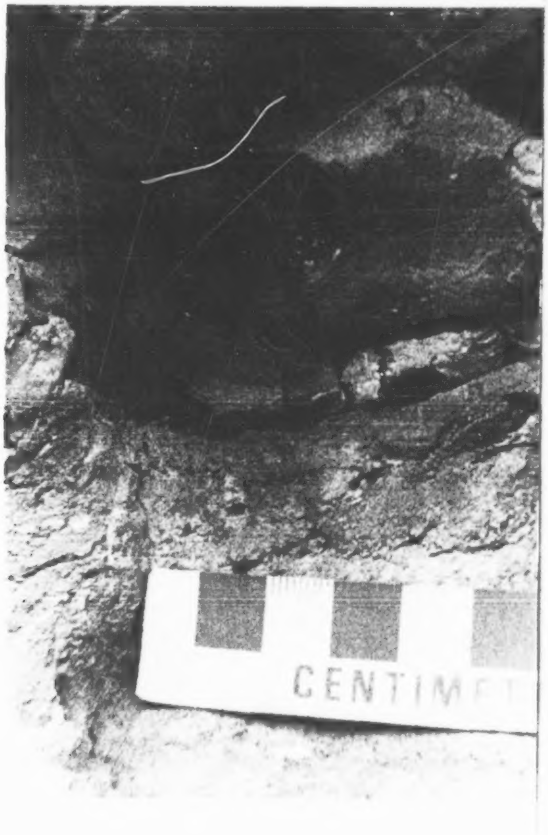


Figure 3-16. h) Massive to weakly laminated muds (facies Fm) deformed by overlying cobble; note vertical fractures interpreted as desiccation cracks.

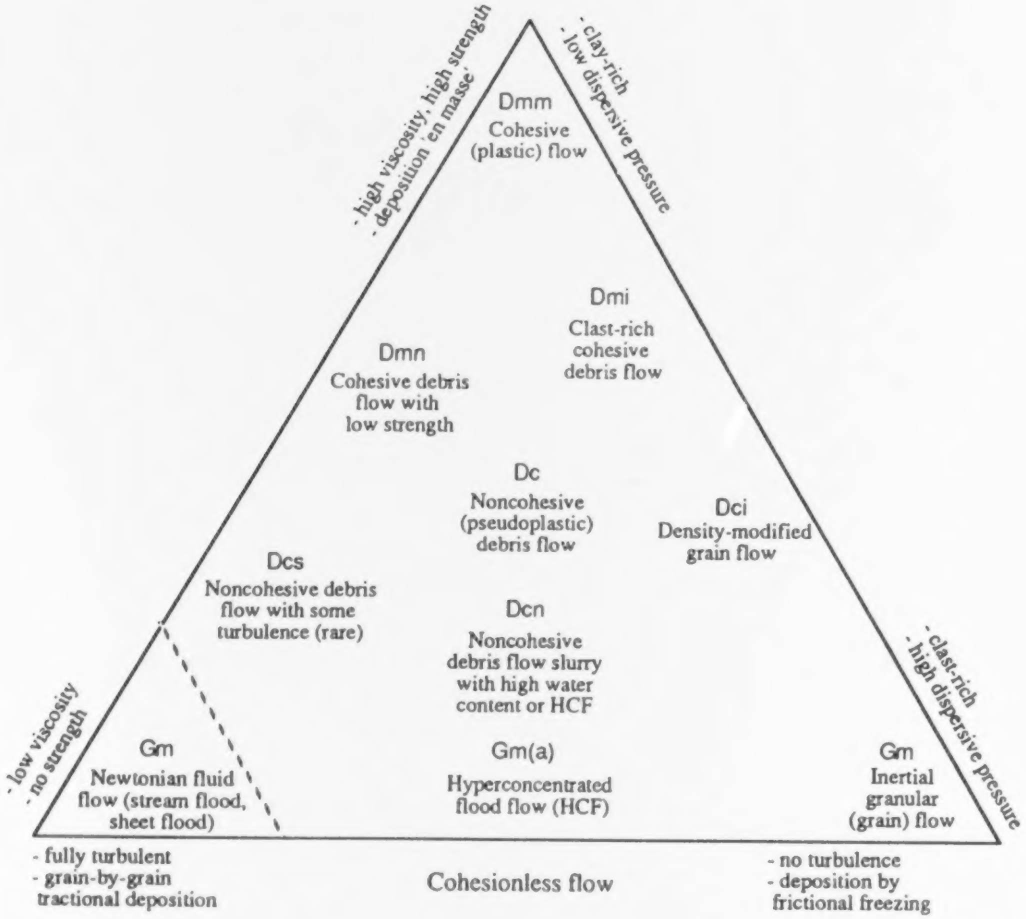

- high viscosity, high strength
- deposition 'en masse'
- clay-rich
- low dispersive pressure
Dmm
Cohesive (plastic) flow
Dmi
Clast-rich cohesive debris flow
Dmn
Cohesive debris flow with low strength
Dc
Noncohesive (pseudoplastic) debris flow
Dci
Density-modified grain flow
Dcs
Noncohesive debris flow with some turbulence (rare)
Dcn
Noncohesive debris flow slurry with high water content or HCF
- low viscosity
- no strength
- clast-rich
- high dispersive pressure
Gm
Newtonian fluid flow (stream flood, sheet flood)
Gm(a)
Hyperconcentrated flood flow (HCF)
Gm
Inertial granular (grain) flow
- fully turbulent
- grain-by-grain tractional deposition
Cohesionless flow
- no turbulence
- deposition by frictional freezing

Figure 3-17. Ternary diagram of subaerial sediment-water flows illustrating Jasper alluvial fan lithofacies and respective interpretations. Rheologic variations are schematically represented as a function of matrix, clast and water content. Matrix-supported, clay-rich flows occur near the top; increasing water content or clast (sand / gravel) content results in flows plotting closer to the lower left or right, respectively.

Location	Area (A_d) (km^2)	Length (km)	Width (km)	Basin slope (°)	L/W Ratio	Basin Relief (H_d) (metres)	Paleofan slope (°)		Ruggedness
							(bedding^3)	(valley^4)	(R) $=H_d A_d^{-0.5}$
Amber Mtn.	4.22 ^{1}	3.0	2.5	18.8	1.2	1020	11 - 14	9.5	0.497
Roche Miette	3.65 ^{2}	4.25	1.25	14.9	3.4	1130	3 - 9	12.5	0.591

[1] horizontally projected area above 1600 m; [2] horizontally projected area above 1200 m; [3] regional dip; [4] valley slope along paleofan reach

Table 3-1. Characteristics of drainage basins above paleofan successions at Amber Mountain and Roche Miette.

Code	Lithofacies - subfacies	Sedimentary Characteristics	Interpretation
Dm	Matrix-supported diamicton: massive; sandy mud matrix	typically ungraded; outsized clasts common; clasts angular to rounded; elongate clasts oriented parallel to flow; bed contacts typically gradational	Cohesive debris flow (CDF): viscous, laminar flow with high yield strength
Dmn	- coarse-tail grading	- basal concentrations of cobbles and boulders	- CDF with clast settling or early turbulence
Dmi	- inversely graded base or - inverse grading	- diffuse laminae in thin basal zone - high clast content	- CDF with basal shear zone below rigid plug - CDF with weak dispersive pressure or inherited grading
Dc	Clast-supported diamicton: massive; muddy to silty sand matrix	typically ungraded; outsized clasts common; clasts mainly angular to subangular; chaotic or weak a[p] a[i] fabric; bed contacts gradational, rarely erosional	Noncohesive (or pseudoplastic) debris flow (NDF) with low matrix strength
Dci	- with inverse grading or - inversely graded base or - inverse to normal grading	- abundant cobbles and boulders - rare stratification in basal zone - may fine-up into Dcn, Gm, Sm or Fl / Fm	Density-modified grain flows with: - high dispersive pressure - basal traction carpet - late turbulence (whaleback bar?)
Dcn	- with normal grading	- rare, coarse-tail grading	- NDF with clast settling in slurry or HCF(?)
Dcs	- with crude stratification	- weak imbrication; clast clusters; basal scours	- NDF with early turbulence and traction
Gm	Gravel: clast supported; massive or crude sub-horizontal bedding; openwork or sandy matrix	a[t] b[i] imbrication; minor sand lenses; clasts mainly subrounded to rounded; bimodal grain-size distribution; outsized clasts rare; some thin, openwork lenses; rare, angular, gravel sheets	traction deposition: longitudinal bar, channel lag and rare, sieve deposits
Gm(a)	- Gravel: clast or matrix supported; massive or weakly stratified; some normal grading	weak imbrication; clasts show both a[t] and a[p] fabric; matrix poorly sorted and polymodal (pebbly sand); outsized clasts common; clasts subangular to rounded; little fine sand or silt; bed contacts gradational or erosional (scoured)	partially turbulent, hyperconcentrated flood flows (HCF)
Gt	Gravel: trough cross-bedded	clast-supported; imbricated; gravels vary from granule to cobble-sized, commonly openwork	channel-fill gravels
Gp	Gravel: planar cross-bedded	same as Gt	transverse bar deposits
St	Sand: trough cross-stratification	medium to coarse sands and gravelly sands; moderately to well sorted	sinuous-crested dunes
Sp	Sand: planar cross-stratification	same as St	straight-crested transverse bars and linguoid bars
Sr	Sand: ripple bedding - all types	fine to coarse-grained, well sorted, sand lenses (2-10 cm thick and 0.2-2 m long); common in the lee of large clasts; lenses taper downstream	ripples: bar top and scour-fill, waning flow deposits; crevasse-splay deposits
Sm	Sand: massive; rare normal grading	fine to coarse-grained sands; rare pebbles	hyperconcentrated flood flow deposits
Sh	Sand: weak horizontal stratification	beds up to 40 cm thick, several metres wide and with diffuse contacts	upper flow regime plane beds
Fl	Very fine sand, silt: fine parallel laminae	very well sorted; minor wavy ripple bedding	suspension deposition: waning flood and overbank deposits
Fm	Clay, silt	massive, some deformed laminae, dessication cracks	mud drapes; ephemeral, restricted pond deposits

Table 3-2. Description and interpretation of lithofacies in the Jasper paleoalluvial fans. Lithofacies and subfacies codes follow those of Miall (1977, 1978, 1983), Eyles et al. (1983), Smith (1987), Walton and Palmer (1988) and Waresback and Turbeville (1990).

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CHAPTER 4 - LATE QUATERNARY STRATIGRAPHY, SEDIMENTOLOGY AND HISTORY OF THE JASPER TOWN SITE AREA, ALBERTA, CANADA[2]

INTRODUCTION

In this chapter, a stratigraphically and sedimentologically complex sequence of Quaternary deposits exposed at Jasper, Alberta (Fig. 4-1a) is described, and a model explaining their origin is presented. The model includes a discussion of two conspicuous landforms in the region. The first is a geomorphic feature referred to here as a kettle chain. It consists of a linear belt of lakes paralleling the Athabasca River east and northeast of Jasper town site (Fig. 4-2). The second landform is a terrace-like feature that flanks the Athabasca River forming a 50 -150 m high scarp directly northwest of Jasper town site (Figs. 4-2 and 4-3a). The development of this 'terrace' is discussed in relation to stratigraphic and sedimentologic evidence and other geomorphic features in the area.

Location and Setting

Jasper town site is situated near the center of Jasper National Park in the Rocky Mountain Main Ranges (Mountjoy, 1981). Relief in the region is high with mountain peaks commonly rising about 1500 m above valley bottoms. The town lies at the junction of two major valleys occupied by the easterly-flowing Miette River and the northwesterly-flowing Athabasca River (Fig. 4-1a). The two rivers join directly south of the town site and flow northeast, across the northwest-southeast regional structural trend of the bedrock (Fig. 4-2). West of the town site, intensely folded shales and sandstones of the Precambrian Miette Group (Price and Mountjoy, 1970) form a wide, bedrock-cored, bench on which Pyramid and Patricia Lakes are situated (Fig. 4-2). Fluvial activity has incised a valley below the level of the benchland (Fig. 4-1b). Jasper town site sits on a terrace within this inset valley at an elevation of about 1060 m asl, approximately 30 m above the Athabasca River. A steep scarp rises 50 - 150 m above the town site to the terrace top. Although the Pyramid Lake benchland is largely bedrock, Quaternary surficial sediments mantle much of the upper surface. Northeast trending drumlins and flutings occur on the benchland (Reimchen, 1976) and are best developed downvalley of northwest-southeasterly trending bedrock ridges (Fig. 4-2). The benchland is relatively flat along its edge forming a distinct terrace-like feature (Fig. 4-3a). The terrace scarp generally is vegetated and locally has a gentle step-like form. A number of deep gullies with disproportionately small drainage areas dissect the Jasper terrace (Fig. 4-2).

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[2] *A version of this chapter has been published: Levson and Rutter, 1989. Canadian Journal of Earth Sciences. v. 26, p. 1325-1342..*

Previous Quaternary Studies

Early investigators (McEvoy, 1901; Kindle, 1929; and Taylor, 1960) related terraces in the Athabasca valley to beach and deltaic deposits of a large glacial lake. The lake is shown on the glacial map of Canada (Prest, 1968) but is no longer recognized (Levson, 1986; Bobrowsky *et al.*, 1987). The surficial geology in Jasper National Park was mapped by Reimchen (1976) and Holland and Coen (1982). Levson and Rutter (1989) investigated the origin of glacial deposits in the Jasper region. Roed (1975) studied the Quaternary geology east of Jasper Park and found evidence for three glacial advances originating in the Jasper area. All are considered to be Wisconsinan (Rutter, 1984). Roed *et al.* (1967) studied the metamorphic rocks of the Athabasca Valley erratics train and concluded that they were carried by ice in the last main glaciation that moved across the Continental Divide into the Miette and Athabasca valleys (Fig. 4-1). The glaciated 'U-shape' of the main passes along the Continental Divide west of Jasper also attest to this movement of ice.

Two ^{14}C dates from the Jasper region effectively bracket the last major glaciation in the area. A date of 29,100 ± 560 years B.P. (GSC 3792) obtained on wood found in gravels underlying till near Jasper town site suggests that ice moved into the region after that time (Levson, 1986; Levson and Rutter, 1986). In the Pocahontas area (Fig. 4-1), gastropods collected from lake sediments were dated at 11,900 ± 120 years B.P. (GSC 3885) thus providing a minimum date for deglaciation of that area (Bobrowsky *et al.*, 1987).

Luckman and Osborn (1979) found that Late Wisconsinan glaciers in the Jasper region had receded to near their present limits by about 10,000 years B.P.. Only minor advances, extending about one to two kilometers beyond present glacier margins, have since occurred (Kearney and Luckman, 1981).

DESCRIPTION OF DEPOSITS

Quaternary sediments in the Jasper town site area are exposed at four sections along the terrace scarp north and west of the town (Sections 831 to 834, Fig. 4-2). The sedimentary sequences at each of these sections are shown as stratigraphic columns in a schematic east-west cross-section of the Jasper terrace (Fig. 4-4).The information obtained from the four main exposures is supplemented by unpublished data from drill logs produced by Alberta Environment, Parks Canada, and private drilling companies (*e.g.* column 83-5 in Fig. 4-4) and from numerous small exposures and pits.

Sedimentary Units

The lowest exposed sediments in the area are gravels and sands (units 1-3, Section 831, Figs. 4-3b and 4-4) that are unconformably overlain by a diamicton (unit 4). The sedimentary sequence at Section 832 is similar and consists of conformable, gravel, sand

and thin silt beds (units 1-4). Cross-beds in stratified units at sections 831 and 832 dip generally downvalley (340° to 100°). A radiocarbon date of 29,100 ± 560 years B.P. (GSC-3792) was obtained from wood in unit 2 at a depth of 17.9 m below the surface (1115 m asl). Unit 5 at Section 832 consists of sand and gravel beds that dip steeply (30-35°) to the north-northeast (20°) (Fig. 4-5). The unit has a gradational lower contact and grades laterally (towards the valley center) into diamicton of unit 6 (Fig. 4-5). The lower contact of the diamicton is planar and abrupt (Fig. 4-3h) and dips at an angle of about 25° towards the valley center (S-SE).

Sand, gravel, and gravelly diamicton beds dominate Section 834 (units 1 to 6). Beds in the lower part of the section (Fig. 4-3c) dip towards the northwest (276° to 315°). Unit 7 is a silty diamicton that grades laterally to the north into silts and fine sands (Fig. 4-4). The section is capped by up to 3 m of poorly exposed, sandy diamicton (unit 8) that thickens towards the valley center (Fig. 4-4).

At the north end of Section 833 (Fig. 4-3d) gravel and sand beds of unit 1 (Fig. 4-4) dip steeply to the southwest, west, north and northeast (235° to 072°). At the southern end (Fig. 4-3e), pebbly gravels (unit 1a) are sharply truncated by poorly sorted, coarse gravels (unit 2). Fine sands occur near the top of much of the section (unit 3, Figs. 4-3d, e) and are locally truncated by cobbly gravels and/or overlying diamicton (units 4 and 5, respectively). The units above the sand are discontinuous and poorly exposed .

The lithology of small to medium pebbles within each of the main sedimentary units in the area is given in Table 4-1.

SEDIMENT ASSOCIATIONS

Sedimentary units in the area can be conveniently grouped into four main sediment associations, each with similar lithologic, textural, and structural characteristics. The vertical and lateral variations in the lithology, geometry, and orientation of the sediments are discussed below. The sediment associations described here are for the interpretation of cross-sections following the approach of Shaw (1987). Interpretations follow the descriptions for each sediment association.

Sediment Association 1 - Stratified Gravels, Sands and Silts

Description

Sediment association 1 consists mainly of horizontally stratified gravels, sands and minor silts. It is represented by units 1 to 3 at Section 831 (Fig. 4-3b) and units 1 to 4 at Section 832 (Fig. 4-5). Trough and planar cross-bedding and scour and fill structures also occur. The grain size distribution and sorting of individual beds are laterally uniform over tens of meters. Fining up sequences dominate and are 3 to 6 m thick except for the 15 m thick sequence at Section 831(units 1 to 3).

Gravel beds occur at the base of the fining upwards sequences and exhibit large scale, crude horizontal bedding and some normal graded bedding. Beds average 40 to 80 cm in thickness. The gravels are imbricated, poorly to well sorted, clast-supported and generally have a medium to coarse sand matrix. Clasts are mainly pebble- to cobble-sized and subrounded to well rounded. Planar cross-stratified sand beds up to 4 cm thick occur sporadically within the gravels.

Clast-supported gravels grade into matrix-supported pebbly sand beds. Horizontal bedding dominates but some planar and trough cross-bedding also occurs. Sand strata dominate the upper part of each cycle and exhibit horizontal laminae and planar and trough cross-bedding. Well sorted fine sands with type A and type AB-transitional climbing ripples (Jopling and Walker, 1968) also are present. Massive sand beds up to one meter thick occur locally. Dish structures are common in some sand beds.

Most cycles are capped by a few centimeters of sandy silt. The silt beds sometimes extend laterally for tens of meters before pinching out. They exhibit wavy laminae, load and injection structures, and may contain up to 50% granules and small pebbles.

Some beds within this sediment association coarsen upwards. For example, unit 4 at Section 832 (Fig. 4-4) coarsens up from fine sands to very coarse sands. Scour and fill structures increase in abundance towards the top of the unit as does the coarseness of the lag gravels at the bottom of the scours. Some overall fining up units may have a thin (up to 0.5 m) coarsening up sequence at the base.

Organic materials, including wood and small gastropod and pelecypod shell fragments, occur in most deposits of this sediment association.

Interpretation

Sediments of this association are interpreted as braided stream outwash deposits. The clast roundness, sorting and stratification suggest that deposition occurred in a fluvial environment. Crude horizontally bedded gravels are typical of coarse grained proximal braided rivers in glacial environments (Church and Gilbert, 1975). The scarcity of trough and planar cross-bedding relative to horizontal bedding, is probably a result of deposition occurring mainly in longitudinal bars as is typical of gravel-bed braided streams (Smith, 1974). Planar stratified, gravel, longitudinal bars 30 to 100 m long dominate the upper Scott and Yana outwash fans (Boothroyd and Ashley, 1975). Horizontally bedded, clast-supported, imbricated gravels are typical of the proximal reaches of the Donjek River (Rust, 1972). Sand and mud facies are also present in these proximal braided stream deposits but they are more common in distal reaches. Rust (1978) found that fine sediment made up 10 to 15% of the upper member of the Cannes de Roche Formation interpreted as distal braided river deposits. Boothroyd and Ashley (1975) observed sand lenses a few tens of centimeters thick and 5 to 20 m long which were deposited during falling stages of flow on the upper Scott and Yana outwash fans. Sand beds with some types A and B ripple-drift cross-laminae commonly overlain by a veneer of silt and clay were common on the middle and lower reaches of the fans.

The fining upwards sequences characteristic of this sediment association probably formed as a result of waning flow conditions subsequent to major flood events. Braided stream deposits characterized by fining-up gravel and sand cycles, formed under progressively decreasing energy levels in successive flood events have been described by several authors (Rust, 1972; Boothroyd and Ashley, 1975; Miall, 1977; and Rust, 1978). Rust (1972) reported fining-up cycles 1 to 20 m thick but indicated that the thicker cycles may represent successive events of vertical aggradation. No disconformities are present in the thick sequence at Section 831 suggesting deposition in either a large flood or even a jokülhlaup event. Re-occupation of the channel during subsequent high flow events resulted in cyclicity of the sedimentation pattern. Preservation of fine-grained sediments at the top of each cycle suggests that subsequent flows were initially depositional and increased in strength gradually. This is supported by a coarsening upwards of grains at the base of some otherwise fining-upwards sequences.

Fine grained deposits do not dominate this sediment association but their presence suggests a more distal environment than that indicated in proximal outwash models discussed by Rust and Koster (1984). Alternatively their presence may be due to more local variations such as deposition in a secondary channel on a higher terrace or along the valley side some distance from the main channel system. Rust (1972) observed that silts and clays, deposited from quiet waters in recently active channels cut-off by falling water levels, formed a thin film of parallel laminated mud. The preservation potential of these muds was low except in more distal or inactive reaches. Silts and clays capping fining-up cycles in this sediment association at Section 832 were probably deposited after temporary channel abandonment, as Rust (1972) observed on the Donjek River. This exposure occurs high along the valley margins in a position where channel abandonment would be likely.

The presence of cut and fill structures in unit 4 (Section 832) suggests some variation from the depositional environment discussed above. The large size of lag gravels at the base of the cut and fill structures indicates that relatively high energy periodic flows occurred. Wide variations in discharge and sediment transport associated with flooding and other causes are typical of proximal proglacial fluvial environments (Church and Gilbert, 1975). The upward increase in the number of cut and fill structures, in the overall grain size of the unit, and in the coarseness of the lag gravels suggests an increased proximity to the glacial source.

Sediment Association 2 - Diamicton with Trough-Shaped Sand and Gravel Lenses

Description

Sediment association 2 consists of massive, matrix-supported diamicton containing abundant trough-shaped (with a planar upper surface and a concave lower surface) sand and gravel lenses (Figs. 4-3b and 4-3f). In the Jasper town site area it is represented only by unit 4 at Section 831 but similar deposits have been observed at other localities in the region (Levson and Rutter, 1989). The sands and gravels are well sorted and exhibit undisturbed large scale trough cross-bedding (channel-fill cross-bedding) with some crude horizontal laminae and weak planar cross-beds. At Section 831, the lower contact of the diamicton (Fig. 4-3f) consists of a series of concave, erosional 'scours' which are similar in size and shape to the sand and gravel lenses within the diamicton (1 to 5 m wide and 0.5 to 1 m thick). A gravel lag occurs at the base of some of the scours (Fig. 4-3f). Small angular inclusions of sand occur near the base of the diamicton. The diamicton has a silty sand to muddy sand matrix. Clasts are mainly subangular to rounded, granules to medium cobbles with rare large cobbles and boulders. The diamicton has a bimodal pebble fabric with a moderately strong preferred orientation (S_1=0.65) trending parallel to the valley (Fig. 4-7a).

Interpretation

Diamictons of this association are interpreted as deposits of sediment flows emanating from the glacier terminus region. Trough-shaped sand and gravel lenses were probably deposited in small subaerial fluvial channels as indicated by the sorting of the deposits, the geometry of the lenses and the presence of undisturbed channel fill cross-bedding. The interbedding of diamicton with these fluvial deposits further suggests that the diamictons were deposited as debris flow deposits. Pebble fabrics characteristic of these diamictons are typical of debris flow deposits (Marcussen, 1975; Lawson, 1979b and 1982). Sand inclusions near the diamicton base are inferred to be rip-up clasts. The 'scoured' contact with lag gravels at the base of the diamicton indicate subaerial exposure. Similar characteristics have been described by Lawson (1979a) in modern glacially derived debris flows. De Jong and Rappol (1983) also interpreted interstratified diamicton and waterlain sediments, with gully-like erosive lower contacts, and highly variable clast orientation and grain size characteristics, as debris flow deposits.

The presence of numerous lenses within the diamicton suggest that it was deposited as a series of debris flows rather than by one single event. Intermittent fluvial activity deposited the sands and gravels that were covered by subsequent flows, thus leaving the stratified lenses within an otherwise massive diamicton. Boulton (1972) found that the coalescence of several debris flows may produce an apparently continuous diamicton with a sharp lower contact. The similarity in size and shape of the sand lenses and the scours at the base of the diamicton(s) and the gravel lag at the base of the scours, suggests a genetic association, with both debris flows and channelized fluvial flows occurring in the same environment.

An alternative origin for the diamicton is that it was deposited subglacially either by meltout, lodgment or flow. Deposits with similar sorting, matrix texture, and clast shape are generally considered to be more typical of debris flows than subglacial tills (Boulton, 1976; Dreimanis, 1976). In addition, characteristics of subglacial tills, such as lodged and heavily striated clasts, shear structures, a strong downvalley fabric, overall low dip of clasts, and compressively deformed sand lenses, are absent. The diamicton may also have originated as a mudflow from the valley wall but the downvalley fabric and the presence of gravels on the slopes directly above the section suggest that this is also unlikely.

Sediment Association 3 - Massive Diamictons

Description

Sediment association 3 consists mainly of massive, matrix-supported diamictons and is represented by unit 6 at Section 832 and unit 8 at Section 834 (Fig. 4-4). Diamictons of this association also are exposed along roadcuts and steep scarps throughout the Pyramid Lake benchlands. Two different vertically associated facies of massive, matrix-supported diamicton can be distinguished in the area. The first is usually the lower, has a silty sand to silty clay matrix with 30-40 % clasts and is very compact. Clasts are mainly subangular to rounded, up to 0.5 m in diameter and locally derived. Faceted and striated clasts are common. The diamicton has an abrupt, planar, erosional lower contact (Fig. 4-3h). On the benchlands this facies occurs commonly below surface deposits and is associated with fluted landforms. Small outcrops of this facies occur about half way up the Jasper terrace above section 831 and, notably, reveal very large clasts (up to 2.5 m in diameter) with strongly faceted and striated upper surfaces. The striations parallel the valley orientation. At Section 832, massive diamicton (unit 6) of this facies grades laterally into sand and gravel beds that dip steeply (30 -35^{o}) towards the north (sediment association 4). Near the transition zone the diamicton exhibits crude stratification with dips and dip directions comparable to those in the adjacent sands and gravels (Fig. 4-5). The lower contact of the diamicton dips toward the valley center (S-SE) at an angle of about 25^{o}.

The second diamicton facies in this association is also massive and matrix-supported. It differs in that the matrix is sandy and less dense, clasts are very angular to subrounded and up to boulder size, and far traveled metamorphic erratics are present. This type of diamicton occurs on the surface at a number of localities on the Pyramid Lake benchland. It is usually poorly exposed due to the loose sandy texture and thinness of the deposits (generally up to about one meter thick) and consequently, detailed descriptions could not be obtained. Locally this facies may be absent or represented only by scattered metamorphic erratics overlying bedrock. These erratics were not observed in diamictons below the surface at any locality. In exposures where both diamicton facies in this association occur, the sandier diamicton with metamorphic erratics is invariably the uppermost. Borehole data from near Patricia Lake, suggest that about one meter of sandy diamicton on the surface overlies a hard, silty diamicton in the subsurface (Column #83-5 on Fig. 4-4). Numerous poorly exposed small sand and gravel pits also occur on the surface of the Pyramid Lake benchland.

Interpretation

The two vertically associated massive diamictons that occur in the region are interpreted as subglacial till overlain by supraglacially derived debris flow deposits. Subglacial tills are generally denser, have stronger, unimodal pebble fabrics oriented parallel to the valley, finer matrix textures, lower total clast contents, and more striated clasts than supraglacially derived diamictons (Boulton, 1976; Dreimanis, 1976; Lawson, 1979b; Shaw, 1982; Haldorsen and Shaw, 1982; Levson and Rutter, 1989).

Diamictons on the Pyramid Lakes benchlands that are massive, clay rich, very compact and hard are interpreted as subglacial tills. The presence of large, apparently lodged clasts (clasts with abundant striations, strongly faceted upper surfaces, and long-axis and striation orientations parallel to the direction of ice flow) suggest that lodgment processes were at least locally involved in the deposition of these tills. The association of these diamictons on the benchlands with streamlined landforms further confirms that they were deposited subglacially.

Similarly, the massive diamicton exposed at Section 832 (unit 6) is interpreted as a subglacial till of either melt-out or lodgment origin. The sharp erosional lower contact, massive structure, dominance of local lithologies, overconsolidation, relatively low clast content and fine matrix texture clearly suggest a subglacial origin. Fractured quartzitic clasts and abundant load and other deformation structures in the underlying sediments may be the result of pressure induced by an overriding glacier. Pebbles within the diamicton exhibit a preferred plunge toward the south (Fig. 4-7b). This fabric may have been formed subglacially as ice expanded up the valley side or, less likely, it may be partially due to post-depositional, gravity-induced movement. The variance between the dominant pebble orientation and the valley trend may also be due to complicated ice flow patterns resulting from the coalescing Miette and Athabasca valley glaciers or may reflect the early northward advance of Athabasca valley ice into the area.

The sandy, less dense matrices, greater clast angularity and larger modal clast sizes of diamictons generally occurring in the upper part of this sediment association, suggests that they were supraglacially derived. Similar characteristics have been described for supraglacial tills by numerous authors (Sharp, 1949; Drake, 1971; Stewart and MacClintock, 1971; Boulton and Eyles, 1979; Lawson, 1979a, 1981a and 1981b; Levson and Rutter, 1989). The presence of distally derived (>50 km) metamorphic erratics on the surface, as well as the general thinness of the diamictons and their common occurrence on the surface, further indicates a supraglacial origin. Many of the metamorphic erratics associated with this facies are soft and easily disaggregated due to their well developed foliation and it is unlikely that they could have survived transport for any significant distance except in a supraglacial or englacial position. The absence of these erratics in diamictons below the surface supports this interpretation. Due to the poor exposure of the diamictons little can be inferred about their mode of deposition which probably occurred in the form of debris flows off the glacier surface or by gradual let-down by melting during deglaciation. The numerous sand and gravel deposits that also occur on the surface of the Pyramid Lake benchland were probably deposited by meltwaters during deglaciation.

Sediment Association 4 - Sand, Gravel, and Gravelly Diamicton

Description

Sediment association 4 consists of well sorted and stratified sands and gravels interbedded with gravelly diamictons. This sediment association is commonly exposed in high valley terraces throughout the lower Athabasca valley in Jasper National Park. In the town site area it is represented by unit 5 at Section 832, units 1 to 7 at Section 834 and units 1 to 5 at Section 833. The grain size distribution and sorting of individual beds are highly variable and change abruptly from bed to bed. Poorly sorted coarse gravels are commonly interbedded with well sorted fine sands (Fig. 4-3c). Lateral and vertical variability over short distances also occurs within beds. For example, pebbly gravels may grade laterally or vertically within 1 or 2 meters into matrix-supported bouldery diamicton (Fig. 4-3g). The dip directions of strata are generally oblique or transverse to the valley commonly indicating paleoflows from the valley center to the valley sides (Fig. 4-3d) with some strata dipping up valley. Steep dips, ranging from 18° to 35°, are common in these areas (*e.g.* Section 833). In other areas (*e.g.* Sections 832 and 834) beds dip more consistently towards the valley side and are shallower with average dips of about 5° (Fig. 4-3c). Normal faults and deformation structures indicating collapse towards the valley center are common (Fig. 4-6a) as are load and injection structures. Beds in the lower part of Section 834 are cross-cut by vertical clay 'dikes' (Fig. 4-6b).

Gravels exhibit horizontal stratification and planar cross-stratification. They range from well to moderately well sorted, pebble gravels to poorly sorted, clast to matrix-supported, large pebble to boulder gravels. Lower bed contacts are typically erosional. Some irregularly shaped silt pockets (up to 30 cm long and 5 cm thick) with weak trough cross-laminae occur in well sorted pebbly gravel beds. Well sorted sand lenses up to 1-2 m long and 5-20 cm thick are also present. Horizontally laminated sands and pebbly sands with normally graded bedding also occur within the gravel dominated units. Poorly to moderately well sorted gravels and sandy gravels grade laterally and vertically into very poorly sorted gravelly diamictons.

Gravelly diamicton units may be either clast-supported or matrix-supported (Fig. 4-3g) although the former dominate. Diamictons generally have a sandy matrix but up to 35% silt and clay may also be present with total gravel clast contents up to 90%. The diamictons exhibit crude, inclined and horizontal planar stratification. Weakly developed scour surfaces are also present. Stratification is best developed in clast-supported beds and generally absent or very weak in matrix-supported units. Beds vary from 0.2 to 3 m thick. Clasts are mainly pebble-to cobble- sized and angular to rounded. Some clasts are striated and are commonly fractured in units overlain by massive diamicton. Matrix-supported diamictons usually are laterally discontinuous and have a silty to gravelly matrix with abundant boulders. Diamicton beds generally have sharp erosional lower contacts especially where they overlie sands (Fig. 4-3g).

Sandy units consist mainly of well sorted medium to coarse sands and very well sorted fine sand and silt beds. Beds are generally two to several centimeters thick. Horizontally laminated sands commonly grade into well developed climbing ripple laminae (types A and B ripple-drift cross-lamination and sinusoidal ripple lamination) (Fig. 4-3c). Trough and planar cross-laminae are also common. Up to 4.5 m of predominately horizontally laminated fine sands with some ripple and trough cross-laminae and pebbly layers are present at Section 833 (unit 3; Fig. 4-4) and angular inclusions of gravel are locally present (Fig. 4-6c). Moderately to poorly sorted pebbly sand beds also occur. Laminated silts and sands are locally interbedded with diamicton (*e.g.* unit 7, Section 834).

The base of this sediment association is observed in only one locality (Section 832). Several centimeters of horizontally laminated silts and sands grade upwards into matrix-supported cobbly sands which become clast-supported within 10 - 20 vertical centimeters (Fig. 4-6d). At Section 832 deposits of this sediment association (unit 5) grade laterally into massive diamicton of sediment association 3 (unit 6).

Interpretation

Deposits of this sediment association are interpreted as lateral, ice-marginal fluvial and debris flow deposits. The structure and sorting of sand and gravel deposits indicate deposition in a fluvial environment. Horizontally stratified and planar cross-stratified, poorly to well sorted coarse gravels are interpreted as high energy fluvial deposits. Normally graded pebbly sands, horizontally stratified coarse sands, trough and planar cross-stratified sands, and planar and ripple laminated fine sands and silts were probably deposited at progressively lower energy levels. The large changes in sediment size and sorting that occur from bed to bed indicate rapid changes in sediment supply and/or stream capacity. The presence of clay dikes and graded beds indicate that sedimentation rates were locally high.

Although the abundance of stratified and well sorted deposits suggests that fluvial activity dominated, the presence of interbedded diamictons indicates the periodic input of unsorted or poorly sorted debris with little subsequent reworking by water. The poor sorting and crude stratification of the diamictons suggests that they were deposited as sediment flows. Variations in the clast content and sorting of the diamictons is typical of sediment flows and is probably due to variable water contents in the sediments during deposition (Lawson, 1979a). The irregular and erosive nature of the lower contact of the diamicton beds may also be explained by a sediment flow origin. Within bed gradations between matrix- and clast-supported gravelly diamictons and poorly to well sorted gravels indicates proximity to a sediment source and the highly variable nature of the flows. The diamictons were probably not deposited by debris flows moving down the valley wall since bedding commonly dips towards the valley side.

Periodic debris flows from glacial ice into an ice-marginal stream would produce sedimentary sequences similar to this sediment association. Large variations in sediment sorting, grain size distribution, clast size and shape, and paleocurrent direction, such as are typical of this sediment association, would be expected in an ice-marginal environment (Boulton and Eyles, 1979). The abundance of climbing ripple laminae in sandy units, suggesting an environment of rapid sediment accumulation, is also typical of ice proximal sediments (Banerjee and McDonald, 1975). The dominance of paleoflow directions transverse to the valley indicate that deposition occurred in a lateral rather than a frontal ice-marginal position. Ice-contact deposits characterized by crudely bedded bouldery gravels with fabrics indicating flow away from the glacier side, have been described at the margins of modern valley glaciers (Boulton and Eyles, 1979). Angular inclusions of gravel in sand beds (Fig. 4-6c) probably are a result of ice-rafting of frozen blocks of previously deposited sediment.

An ice-contact origin for deposits of this sediment association is also indicated by the steep dip of beds (30-35°) away from the valley center and by the lateral gradation of these deposits into massive diamicton interpreted as basal till (*e.g.* Section 832). The lateral facies change is gradational, suggesting that the two deposits originated in adjacent depositional environments. The sediments in the transition zone are texturally similar to the massive diamicton but exhibit a crude stratification with a dip and dip direction similar to the adjacent sands and gravels (Fig. 4-5). Thus, it appears that a glacier occupied the valley center and deposited material at its base. This basal zone graded laterally into an ice-contact terrace along the northern flank of the glacier where the diamicton, sand and gravel of this sediment association were deposited. Boulton and Eyles (1979) and Eyles (1979) described similar depositional environments where supraglacial debris accumulated as scree along steep ice margins while till was deposited at the glacier base.

This interpretation is further supported by the presence of high angle normal faults in deposits of this sediment association. Most faults dip towards the valley center and probably formed when supporting ice in the adjacent valley center melted causing collapse of the sediments. Schwan and Van Loon (1979) also found widespread normal faulting associated with recession of the ice "shoulder" supporting kame terrace deposits.

The base of this sediment association exposed at Section 832 consists of a matrix-supported cobble bed supported by, and overlying, laminated fine sands. The cobbles must have been transported by a relatively high energy medium which presumably, should have deeply eroded the sands. Laminae within the sands, however, are relatively undisturbed by the cobbles and the contact between the sands and cobbles is gradational (Fig. 4-6d). This suggests that the cobbles were gravitationally 'dumped' into the stream depositing the sands, possibly off an adjacent ice surface. Sharp (1949) described similar depositional environments on the Steele Glacier where well sorted cobble deposits occurred on ice surfaces slightly above adjacent supraglacial streams.

Horizontally laminated silts and sands in the upper part of this sediment association at Section 834 indicate quiet water sedimentation along the ice margin. The silts are laterally interbedded with thin diamictons interpreted as small debris flows into an ice-marginal pond.

DISCUSSION

Stratigraphy

The lateral and vertical stratigraphic relationships of Sections 831 to 834 are presented schematically in Figure 4-4. The main stratigraphic units are separated by dashed lines.

Lower Gravels and Sands

Units 1 to 3 at Section 831 are interpreted as belonging to the same stratigraphic unit as units 1 to 4 at Section 832 (Fig. 4-4). At both sections horizontally bedded gravels and sands of sediment association 1, interpreted as braided stream deposits, are unconformably overlain by glacial diamictons (sediment associations 2 and 3). At Section 831 the gravels and sands fine upwards, which may reflect a decrease in stream energy subsequent to a major flood event. The topographically high position of the gravels at Section 832, their strongly cyclic nature and the scour and fill structures lined with coarse clasts in unit 4 suggest that at least their upper parts were deposited in a more proximal, ice-marginal, position. The coarsening upwards cycle at the top of the sequence (unit 4) may be due to the increased proximity of glacial ice. This is further indicated by the gradational contact of unit 4 with the overlying ice contact deposits (unit 5) discussed below. Organic materials (mainly wood and shell fragments) were obtained from the upper part of the gravels at Section 831, and the lower part of the gravels at Section 832, also suggesting that these gravels were deposited in a relatively distal environment compared to the uppermost gravels at Section 832 (unit 4).

The gravels encountered in borehole 83-5 are probably similar to those of sediment association 1 at Sections 831 and 832. These gravels would have been deposited when ice levels were high enough to allow for the development of proglacial streams on the Pyramid Lake benchland.

Diamicton Units

Several types of glacial diamicton overlie the gravel sequences. Unit 4 at Section 831 consists of diamicton with trough shaped sand and gravel lenses (sediment association 2), interpreted as proglacial debris flow sediments with intermittent fluvial deposits. The diamicton occurs over the most distally deposited units of the lower gravels and sands. At Section 832, the more proximal gravels are overlain by a massive diamicton (unit 6) which is interpreted as a subglacial till. The two diamicton units are not exposed together in any of the town site sections but the small exposures of subglacial till above Section 831 and the relative topographic position of unit 6 at Section 832 and unit 4 at Section 831, shows that diamictons interpreted as subglacial tills overlie those believed to be proglacial deposits. The vertical sequences at Sections 831 and 832 may thus indicate a progression from distal to more proximal gravels and from proglacial debris flows to basal tills. These relationships are probably a record of the advance of a glacier into the area and its expansion up the valley sides. Deposition along the lateral margins of this glacier is indicated by the lateral change at Section 832 from subglacial till (unit 6) to lateral ice contact deposits (unit 5).

This interpretation is confirmed by lithologic data on the gravels. Section 831 lies within the Athabasca River valley while Section 832 is within the Miette River valley.

Significant carbonate outcrops occur only in the drainage basin of the Athabasca River. The drainage basin of the Miette River lies entirely within the Main Ranges where clastic rocks predominate. Limestone and dolostone clasts are common at Section 831 but rare or absent in the lower part of Section 832 as expected (Table 4-1). Some carbonates, however, are present in the uppermost gravels (unit 5) at Section 832, suggesting a change from the original Miette Valley source area. The new input of carbonate clasts into the Miette River system in the vicinity of Section 832 probably was due to the movement of Athabasca ice into the area. Thus the increase in carbonates at the top of Section 832 may also be explained by the advance of ice into the area. Athabasca valley ice must have arrived in the area first in order for subaerial sediments to be deposited in the Miette valley at the margin of the Athabasca valley glacier.

Surface diamictons of sediment association 3, which occur throughout the Pyramid Lake benchlands, are correlative with the surface diamicton at Section 832 (unit 6). The diamictons on the benchlands are interpreted as a basal/supraglacial till sequence and indicate that ice overrode all of the deposits discussed above. The fluted and drumlinoid topography of the benchland area contrasts with that east of Jasper town (Fig. 4-2). Although flutings and drumlins may form by subglacial erosion or deposition either by ice or by meltwater, their presence clearly indicates an overlying glacier. The local presence of large scale erosional features typical of flutings and drumlins formed as a result of subglacial meltwater activity (Shaw and Sharpe, 1987) indicates that a similar origin is possible for at least some of these features. The origin of these streamlined forms in the Jasper area was not investigated in detail, but shallow pits indicate that they are mainly composed of diamicton with a fine grained matrix. The presence of only isolated supraglacial deposits and scattered surface erratics on the benchlands suggests that large volumes of supraglacial debris did not cover the ice there. The elevation of the ice in the Jasper town site region was about 2240 m asl, as indicated by the maximum elevation of metamorphic erratics found in the western slopes of Whistlers Mountain (mainly muscovite-garnet schist and quartz-biotite gneiss). Bimodal fabrics such as that present in the upper diamicton at Section 831 (Fig. 4-7a) may in part be the result of reorientation of clasts by overriding ice (MacClintock and Dreimanis, 1964; Ramsden and Westgate, 1971).

The basal/supraglacial sediment association common on the Pyramid Lake benchlands is not well represented along the margins of the Jasper terraces. Section 834, for example, is capped by a thin, matrix-supported, massive diamicton of unknown origin (unit 8). The diamicton and its lower contact are poorly exposed but the presence of metamorphic erratics on the surface suggest that it is, at least in part, supraglacially derived.

Deposits of Sediment Association 4

Sections 833 and 834 are dominated by deposits of sediment association 4, interpreted as lateral ice-marginal sediments. The lower contact at both sections is covered and consequently the nature of the underlying sediments and contacts with them are unknown. The sediments at both sections may lie either stratigraphically below or above the diamicton interpreted as a subglacial till at Section 832 and thus they may have been deposited either prior to or after, respectively, the main advance of ice into the region.

There is evidence that some or all of the sediments at Section 834 were deposited prior to the advance of ice over the area. The facies changes at Section 832, from pre-advance outwash deposits to ice-marginal sediments to subglacial till, indicate that some ice-marginal sediments in the region were deposited during the advance stage. The similar elevation and proximity of the ice-marginal sediments at Sections 834, to those at Section 832, suggest that they were also deposited during the advance of ice into the area. The increase in mean clast size and frequency and thickness of gravelly diamicton beds towards the top of Section 834 (Fig. 4-4) may indicate the gradual advance of a glacier. In addition, load and injection features, possibly induced by an overriding glacier, are present at both sections. High angle faults are typical of sediments deposited on the margins of ablating glaciers. Their absence at Section 834, further suggests that the sediments were deposited during the advance phase. The occurrence of flutings on the adjacent terrain also supports the interpretation that ice overrode the entire sequence of deposits at Section 834.

Pebble lithology data from Section 834 (Table 4-1) show that, in addition to the presence of metamorphic erratics, few carbonates are present in the surface diamicton(s) (unit 8) in comparison to underlying units (*e.g.* Sample 834-6) or deposits of sediment association 4 elsewhere in the region (*e.g.* Samples 832-5, 833-1, and 833-4). This indicates that the diamicton is derived from the Miette valley to the west with little or no input from the Athabasca valley. In contrast, the deposits of sediment association 4 were presumably deposited at the margins of an Athabasca valley glacier. These data suggest that unit 8 was not deposited as a sediment flow in association with the underlying ice-marginal deposits but that it may have been deposited by an overriding Miette valley glacier.

The thin surface diamicton(s) at Section 834 may represent the only remaining depositional evidence of the overriding glacier. Although their thinness (< 1 m) suggests that this is unlikely, surface tills on the adjacent fluted terrain are locally very thin or absent (due to either non-deposition or postglacial erosion). In some areas supraglacially transported metamorphic erratics lying directly on bedrock or on thin basal tills are the only evidence of the entire basal/supraglacial sequence deposited during the last major glacial episode in the region.

Similarly, the surface diamicton at Section 833 (unit 5) may be either a subglacial till, a supraglacially derived debris flow deposited during ice retreat from the area or a postglacial colluvial deposit. If either of these latter possibilities is correct, then part or all of the deposits underlying the surface diamicton at this section may have been deposited along the margin of a retreating ice mass. These deposits show few load structures and no evidence of shear as would be expected if they were overridden by ice. In addition, the diamicton is discontinuous and no linear flute marks are at its base. The incorporation of underlying sand and gravel deposits, an irregular lower contact and the lateral discontinuity of the diamicton are suggestive of a debris flow origin. The lower contact locally dips toward the valley side suggesting that the diamicton was not derived as a debris flow off the valley wall but, like the underlying deposits, may have had a glacial source in the valley center

All of the deposits at Section 833 are therefore interpreted as supraglacially derived debris that flowed and washed off of a stagnating ice mass. The dominant dip direction of bedding, which indicates paleoflow directions towards the valley side, and the location of the section on the side of a pre-existing bedrock valley, support this interpretation. The great thickness of deposits interpreted as ice-contact sediments at Section 833, requires a large sediment source on the ice and it seems likely that there would be more supraglacial sediment during ice retreat than during the advance stage. In addition, the presence of normal faults with large displacements indicate that substantial collapse has occurred probably as a result of melting of supporting ice. The sediments probably accumulated in a lateral ice-marginal position, probably in a kame terrace or similar feature.

Thus, the ice-marginal sediments exposed in the Jasper area at Section 832 were clearly deposited during an advance stage. The sediments at Section 833 were probably deposited during retreat of ice in the area. Evidence regarding a post or pre-advance origin for the deposits at Section 834 is inconclusive, but investigation suggests that they were from the advance stage.

Summary

The character and distribution of stratigraphic units suggests the following sequence of events (Fig. 4-8). Horizontally stratified sands and gravels (sediment association 1) were initially deposited in a braided stream in the Athabasca valley (*e.g.* units 1 -3, Section 831). Deposition of sediment association 2 (*e.g.* unit 4, Section 831) began with the first input of debris flow sediments from glaciers up the Athabasca valley (Fig. 4-8a). With the initial movement of ice along the valley, streams near the mouth of the Miette River valley were confined to the valley side and sands and gravels (*e.g.* unit 4, Section 832) were deposited at relatively high topographic positions (Fig. 4-8b). Massive diamicton (*e.g.* unit 6, Section 832) was deposited at the glacier base as ice advanced into the area and began to expand up the valley sides (Fig. 4-8c). At the same time, sand, gravel and gravelly diamicton of sediment association 4 (*e.g.* unit 5, Section 832) were deposited at the ice margin by debris flows and water flowing off, and along the side of, the Athabasca valley glacier. Ice from both the Miette and Athabasca valleys eventually overrode the entire area forming a near horizontal landscape at about the level of the present upper surface of the Jasper terrace (Fig. 4-3a). A supraglacial/subglacial sediment association was deposited by these glaciers on the Pyramid Lake benchland (*e.g.* unit 8, Section 834). During deglaciation (Fig. 4-8d), ice-marginal sedimentation dominated in what is now the town site area, resulting in the formation of a sedimentologically and stratigraphically complex sequence of deposits (*e.g.* units 1 - 5, Section 833).

Geomorphic Evidence

Several geomorphic features in the Jasper area provide further clues to unraveling the glacial history of the region. East and northeast of Jasper town site, a chain of several lakes occurs on the floodplain of the Athabasca River (Fig. 4-2). Some, such as Twin

Lakes and Trefoil Lakes, are small, deep, steep-sided, and circular. The larger ones, such as Edith and Annette Lakes, are composed of multiple basins each of which is similar in size and shape to the smaller lakes. The surficial materials surrounding the lakes generally are coarse textured with numerous cobbles and boulders present. Although rounded clasts dominate, some large angular clasts are also present. The gravels are in places over 30 m thick. The lakes all lie on glaciofluvial terraces (Reimchen, 1976; Holland and Coen, 1982). Although terraced, the topography around all the lakes except Twin Lakes is hummocky (Reimchen, 1976). The hummocks have gentle slopes and are locally obscure (Fig. 4-2). They are best developed around Edith and Annette Lakes.

The size, shape, and linear trend of the lakes suggests that they are either karst sinkholes, thermokarst lakes formed by meltout of ground ice or kettles formed by meltout of glacial ice. The coarse texture of the surrounding surficial deposits would inhibit the formation of massive ground ice required for the development of thermokarst lakes.

An analysis of the local bedrock geology suggests that the lakes are not karst sinkholes. The southwesterly dipping Pyramid thrust lies northeast of the town site in the vicinity of Edith Lake (Fig. 4-4, inset). The thrust separates carbonates, of the Devonian Palliser Formation and the Mississippian Rundle Group, to the northeast from Miette Group sandstone, shale and siltstone to the southwest (Mountjoy and Price, 1985). Annette, Trefoil, Mildred, Beauvert and Twin lakes all lie southwest of the Pyramid thrust. Consequently, they overlie clastic rocks that are not subject to dissolution and the formation of sinkholes and the lake chain must have another origin.

Thus, it can be tentatively deduced that the lakes originally formed as kettles by the meltout of glacier ice partially or totally buried in outwash and/or glacial debris. The small sizes, and deep steep-sided pothole shapes of the lakes are compatible with a kettle origin. The nature of the associated outwash deposits and the surrounding hummocky topography are also suggestive of a kettle origin. Lines of lakes oriented along pre-existing topographic lows have long been considered to be kettles (*e.g.* Thwaites, 1935). A kettle origin for the lakes indicates that a linear mass(es) of stagnating ice at one time occupied the valley east of Jasper town.

Jasper town site lies at the junction of two major valleys and it seems reasonable to assume that a large medial moraine developed in the valley center during the last glaciation of the area. Medial moraines form by the coalescence of lateral moraines and bed-parallel debris septa (Boulton and Eyles, 1979) at the junctions of valley glaciers. Material accumulates on the ice surface both by debris falls from the adjacent mountain sides and by meltout of englacial debris (Washburn, 1935; Sharp, 1949; Boulton and Eyles, 1979). In the final stages of ablation, debris melting out from near the glacier base may also accumulate on the ice surface. Consequently, in addition to supraglacially derived coarse angular debris, more rounded and abraded clasts may also be present (Gomez and Small, 1985). During deglaciation, this relatively thick supraglacial debris cover may then result in the stagnation of the underlying ice and the development of knob and kettle topography (Eyles, 1979).

Thus, both the regional setting and the presence of a chain of kettle lakes in the valley center suggest that a linear belt of stagnating ice may have occupied the area east of Jasper town site. This is further supported by sedimentologic data from deposits exposed above Jasper town site (*e.g.* Section 833). These data indicate that deposition occurred in a kame terrace or similar feature along the western edge of an ice margin with regional paleoflow directions indicating a source in the valley center to the east. The location of Twin, Beauvert, Mildred, Trefoil, Annette and Edith Lakes near the valley center directly downvalley from the junction of two major valleys, the linear trend of the lakes, and their northeast-southwest orientation all coincide with the presumed location, trend and orientation of a former medial moraine between the Athabasca and Miette valley glaciers. This observation implies a possible association between the lakes and the former moraine and its underlying debris septum.

The following sequence of events is postulated to explain the formation of the lake chain. During the last glaciation in the area, a large medial moraine and debris septum formed at the confluence of glaciers moving down the Athabasca and Miette river valleys (Fig. 4-9a). During ice retreat, debris derived from the moraine and from meltout of the underlying debris septum accumulated on the ablating ice surface. The relative proportion of debris from these two sources is not known but, considering the generally low quantity of debris in or below medial moraines, the dominant source of material may have been from the medial debris septum near the base of the glacier. The accumulating supraglacial cover probably was also supplemented to a lesser degree by debris from other medial moraines originating further up the Athabasca and Miette River valleys. The debris insulated the underlying glacier and developed into an elongated ridge of debris covered stagnating ice, characterized by ice-cored debris knobs and kettle lakes (Fig. 4-9b).

When the adjacent terrain became ice-free, debris washing and flowing off the stagnating glacier in the valley center was deposited along the ice margins (Fig. 4-9b). Glaciofluvial reworking of the morainal debris resulted in partial or complete burial of blocks of ice by outwash gravels and the deposition of coarse, terraced, outwash deposits within and adjacent to the stagnating ice terrain (Figs. 4-9b, c). Jasper town site and Twin Lakes are now located on one of these more prominent terraces. In the valley center, continued melting resulted in a linear belt of low lying knob and kettle topography. Further glaciofluvial activity locally obscured the hummocky topography and final melting of underlying ice blocks resulted in a remnant kettle chain in the position formerly occupied by the medial moraine between the Athabasca and Miette valley glaciers (Fig. 4-9c). Subsequent fluvial erosion was largely confined to adjacent areas allowing for the preservation of some of the lakes and the locally hummocky surface of the terrace on which Edith, Annette, Trefoil, Mildred and Beauvert Lakes are now located. Variations in the shape of the lakes and the nature of the surrounding terrain can simply be explained by differences in the amount of glaciofluvial and fluvial reworking of the morainal deposits prior to final melting of the ice blocks (Fig. 4-9c).

A number of other geomorphic features present in the Jasper area also suggest that ice with a substantial supraglacial debris cover stagnated in the valley east of the town site. These features include: 1) kame terraces composed of ice-marginal sediments such as those exposed at Section 833; 2) lateral terraces or benches on the valley wall possibly formed during halts in the glacier's recession down the valley wall; and 3) large gullies along the margins of the terraces (Fig. 4-2) which have small drainage areas indicating that their formation was aided by either glacial meltwater or the melting of buried ice. Boulton and Eyles (1979) described similar features along the pronounced flanks of ice-marginal dump moraines formed during recession of valley glaciers. In the Jasper town site area these geomorphic features formed along the ice margin near the valley center and not along the main valley wall. The presence of a pre-existing valley incised into the bedrock of the main valley and the formation of a medial moraine and debris septum allowed for the development of these geomorphic features near the valley center in the Jasper area. The geometry of this inset valley was presumably similar to that of the modern valley occupied by the Athabasca River (Fig. 4-1b).

Distal proglacial sediments exposed at the base of Sections 831 and 832 must have infilled the Athabasca River valley up to an elevation of approximately 1120 m (Fig. 4-4), prior to the advance of ice into the area (unless the valley center was already blocked by a small glacier, an unlikely event at 29 ka). The kettle lakes in the valley centre indicate an ice depth at least 100 m below this elevation. The relatively narrow and deep incision of these sediments suggests they were eroded by fluvial rather than glacial erosion. A relatively low concentration of rounded quartzite clasts in the tills, as would be expected if ice was responsible for erosion of the underlying gravels, supports this conclusion. In addition, the presence of distal braided stream gravels (Units 1 to 3 at Sections 831 and 832) as well as proximal glaciofluvial gravels and ice-marginal deposits under till (at Sections 831, 832 and 834) indicates that the glacier was not highly erosive in this area. Since proglacial streams are invariably aggradational, the deep erosion of proglacial sediments in the valley center was probably a result of subglacial fluvial erosion. The resultant low area was then occupied by the overlying glacier. Subglacial fluvial erosional features of a similar or larger scale have been recently well documented in the literature (*e.g.* Shaw and Sharpe, 1987; Shaw *et. al.*, 1989; Shaw and Gilbert, 1990; Rains *et. al.*, 1993).

The lake chain extending from Twin Lakes to Edith Lake does not occur exactly half way between the sides of Pyramid Mountain and Mt. Tekarra but lies slightly to the east near the center of the inset valley incised by the Athabasca River (Fig. 4-1a). If the hypothesis that the lake chain reflects the former position of a medial moraine is correct, then ice moving down the Miette River valley must have dominated over Athabasca ice resulting in a slight downvalley deflection of the moraine to the east. This was probably the case as ice in the Miette valley was largely supplemented by ice from west of the Continental Divide. The underlying topography may also have had an influence on the position of the moraine as suggested by the location of the lakes very near the center of the inset valley. In addition, fluvial erosion in the present location of the Athabasca River may have removed evidence of part of a moraine located just further west.

Although the evidence for all of the events in the hypothetical sequence outlined above is sometimes subtle, all of the existing geomorphic and sedimentologic observations are explained by this interpretation. Specifically, these observations include: 1) the presence of hummocky terrain and pitted outwash in the valley center; 2) the presence of a kame terrace and associated geomorphic features along the margin of the inset valley; 3) the presence of exposed sediments along the inset valley wall indicating paleoflows originating in, and post depositional collapse towards, the valley center; 4) the location of Twin, Beauvert, Mildred, Trefoil, Annette and Edith Lakes near the valley center directly downvalley from the junction of two major valleys; 5) the linear trend of the lakes; 6) the northeast-southwest orientation of the lakes; and 7) the kettle-like size and shape of the lakes.

A similar spatial distribution of landforms to that found in the Jasper town site area may occur in analogous geomorphic settings at major valley junctions. In addition, kettle chains of a similar origin could develop along main valley walls in the former positions of lateral moraines. Kettle chains formed in former lateral and medial moraine positions may be more common features than previously recognized.

Glacial History

The advance of ice into the Jasper area is first recorded by the presence of outwash sediments overlain by proglacial debris flow deposits in the valley bottom and confined outwash deposits on the valley sides (Fig. 4-8). Proximal debris flows and ice-contact, stratified sands and gravels were deposited at the glacier margin as it expanded up the valley side. Athabasca Valley ice advanced into the area prior to the domination of glaciers from the west and subglacial tills were deposited over most of the valley bottom when it was fully occupied by ice (Fig. 4-9). During full glacial times, there was little ice erosion of the pre-existing deposits that occupied the Athabasca Valley. The erosive base of the ice apparently established itself near the level of the bedrock benchlands now supporting Pyramid and Patricia lakes; deeper erosion of pre-existing sediments along a narrow zone in the valley center, east of the present site of Jasper town, probably was accomplished by subglacial fluvial erosion. Glaciers moving down the Athabasca and Miette river valleys coalesced and formed a large medial moraine and debris septum in this area.

Ice retreat on the Pyramid Lakes benchlands left a fluted topography with only scattered supraglacial deposits. In the valley center, however, a linear block of stagnating ice developed as a result of the accumulation of debris on the ice surface (Fig. 4-9). The debris was derived from an ablating medial moraine and a medial debris septum extending down the Athabasca valley center from southwest of Twin Lakes to northeast of Edith Lake. Debris washing and slumping down the sides of this stagnating ice mass onto the adjacent ice free terrain was deposited as a small kame terrace west of Jasper town site. During the late stages of retreat, large amounts of material were deposited in these ice-marginal positions. These sediments are characterized by steeply dipping strata deposited under highly variable flow conditions. Small lateral benches were formed as ice ablated down to progressively lower levels. Gullying, faulting and deformation occurred in the ice marginal sediments as underlying ice blocks melted out. The knob and kettle topography which finally developed in the valley center was obscured by meltwater from the receding glacier and by postglacial fluvial erosion. A chain of kettle lakes are remaining evidence of the possible existence of the moraine.

CONCLUSIONS

1) Exposures of Quaternary sediments in the Jasper town site area record a complete cycle of glaciation and deglaciation during the Late Wisconsinan.

2) The pre-existing topography of the town site area had a strong influence on sedimentation patterns during both the advance and retreat stages. Most sedimentation, in the bedrock low in the valley center, occurred in proglacial braided streams and in supraglacial and ice-marginal (kame terrace) environments. Consequently, glaciofluvial materials dominate exposures in the area. On the higher Pyramid Lake benchlands subglacial deposition dominated.

3) The development of a thick supraglacial cover derived from medial moraines and debris septa at the confluence of large valley glaciers may provide a substantial debris source for the formation of ice-contact deposits and knob and kettle topography.

4) Kettle chains and belts of hummocky topography, formed in former medial and lateral moraine positions, may be relatively common geomorphic features in glaciated mountain valleys.

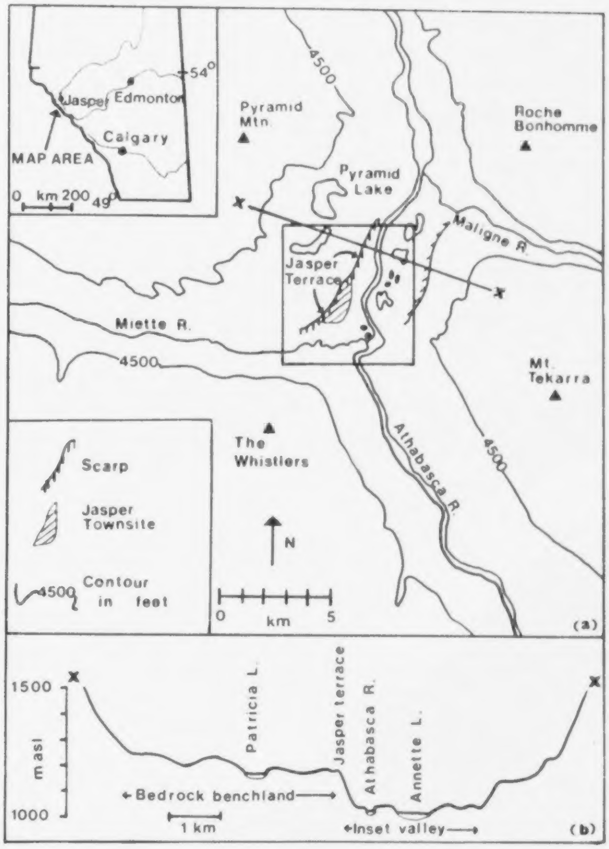

Jasper
Edmonton
54°
Calgary
MAP AREA
0 km 200
49°
4500
Pyramid Mtn.
Roche Bonhomme
Pyramid Lake
Maligne R.
Jasper Terrace
Miette R.
4500
Mt. Tekarra
The Whistlers
Athabasca R.
4500
Scarp
Jasper Townsite
4500 Contour in feet
N
0 km 5
(a)
1500
masl
1000
Patricia L.
Jasper terrace
Athabasca R.
Annette L.
Bedrock benchland
1 km
Inset valley
(b)
X
X

Fig. 4-1. a) Map of Jasper townsite area showing the junction of the Athabasca and Miette River valleys. Note the line of lakes extending directly downvalley from the confluence of the two rivers, and the flanking scarps. b) Cross-section showing Athabasca River valley incised below the Patricia and Pyramid Lakes benchland.

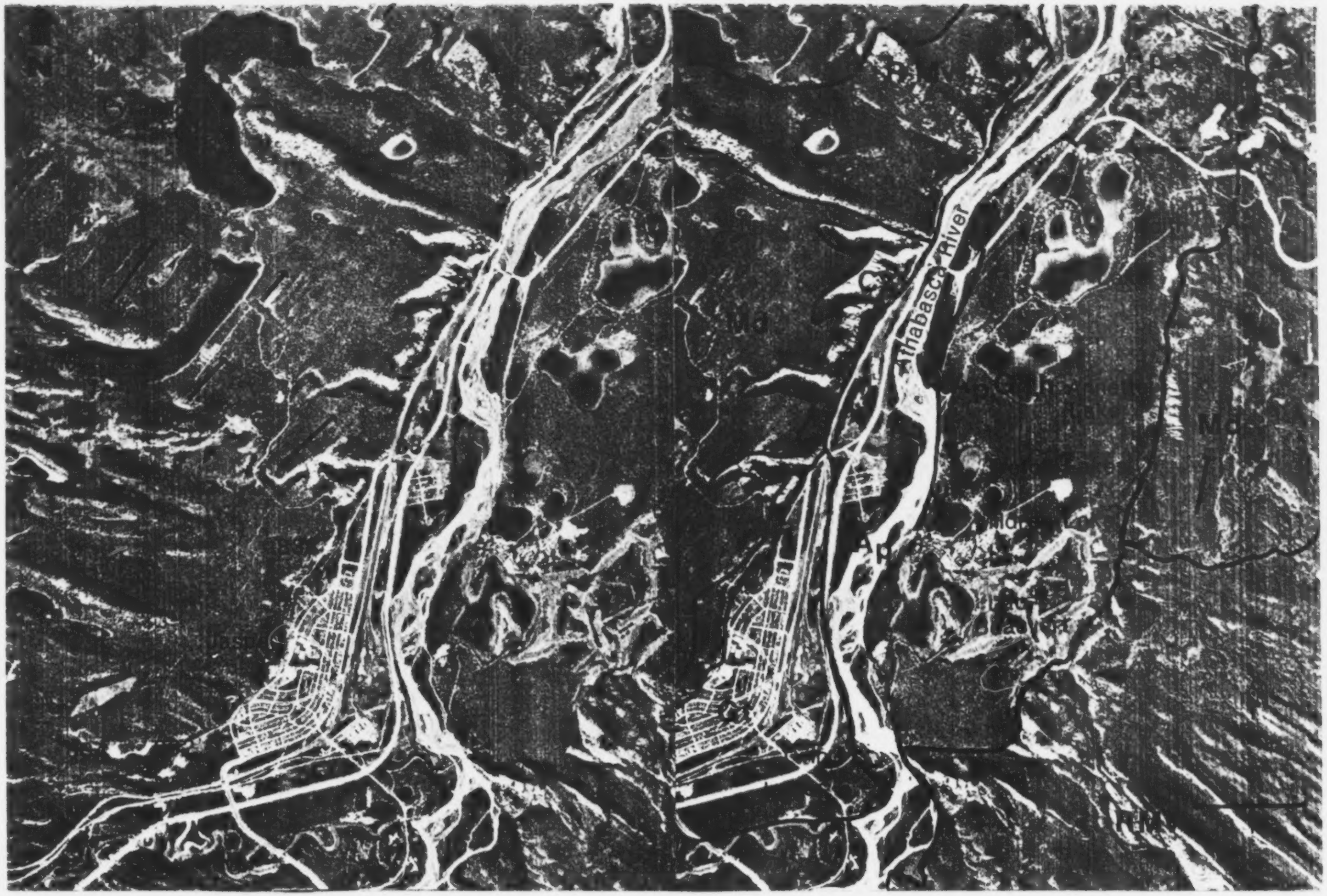


Fig. 4-2 Surficial deposits in Jasper townsite area. Note: linear trend of lakes east of Jasper cross-cutting the structural trend in the area, incised valleys of the Athabasca and Miette rivers, prominent terrace west of Jasper, and fluted terrain in the Patricia Lake region (arrows indicate former direction of ice flow). A, alluvium; C, colluvium; G, glaciofluvial deposits; M, ground moraine; R, bedrock; d, drumlinized or fluted; h, hummocky; p, floodplain; t, terraced; v, veneer

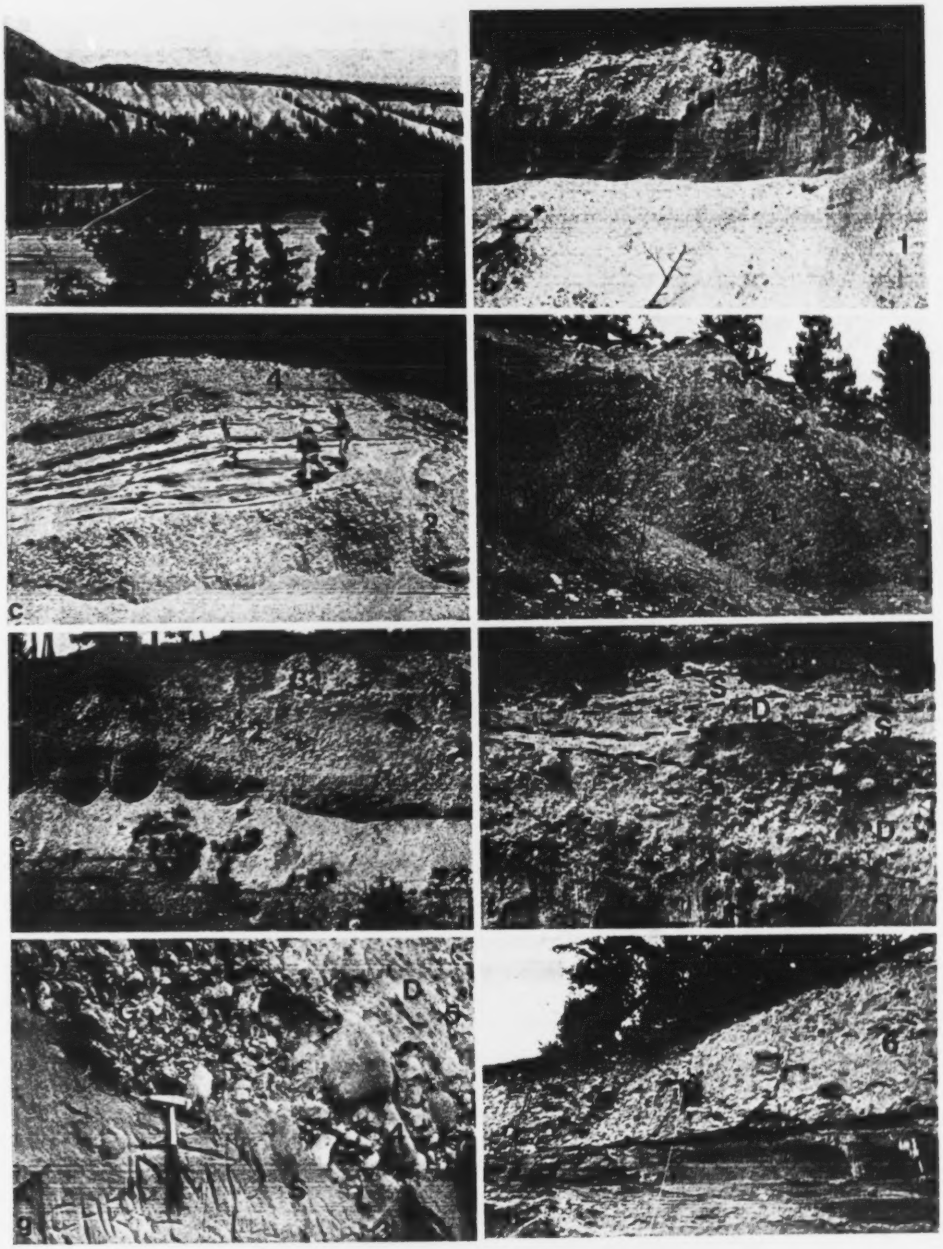
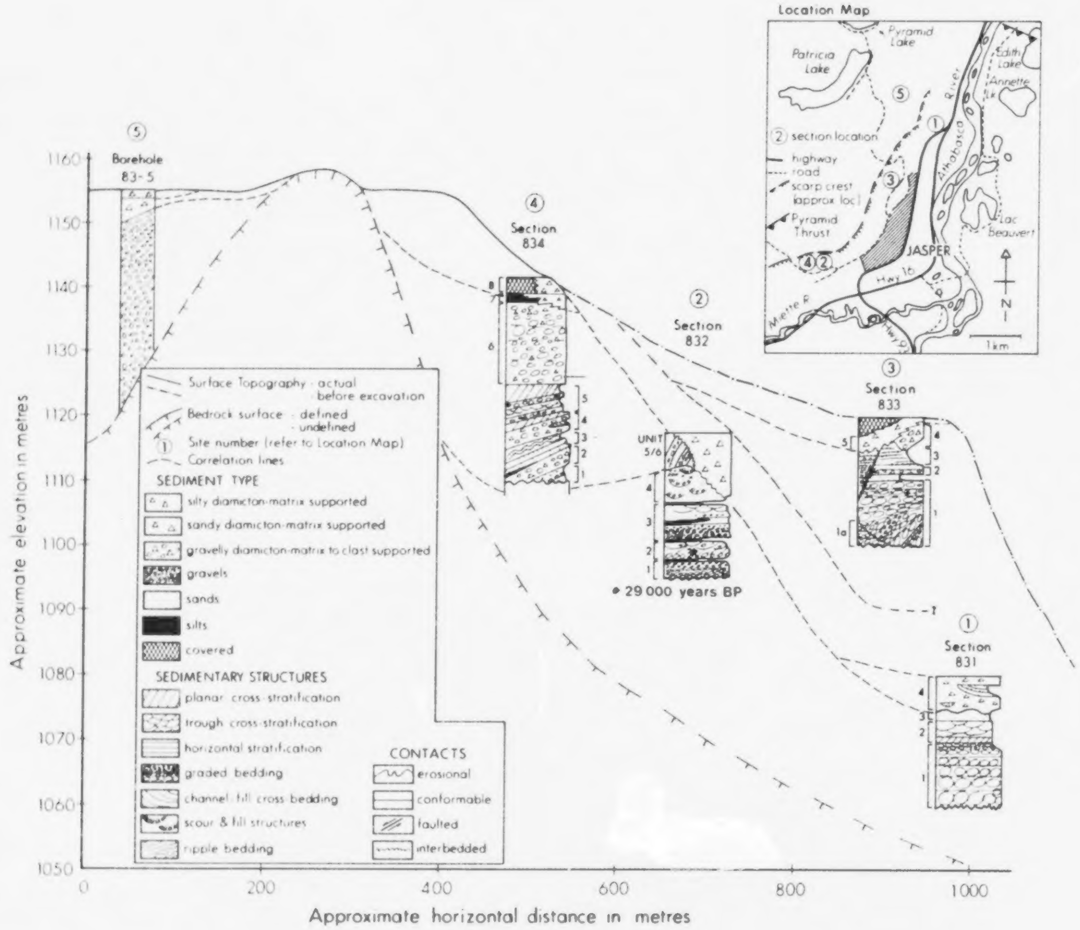


Fig. 4-3 a) Typical terrace-like landform flanking the Athabasca River. b) Section 831 (19.8 m thick). c) Base of of Section 834 showing interbedded, gravel, gravelly diamicton and sands. Beds dip towards the valley side (310-315^{O}). d) North end of Section 833. Unit 2 is absent here. Beds in unit 1 dip 26-28^{O} to the SW (210-235^{O}), valley center to right of photo; exposure 12 m high. e) South end of Section 833 (about 8 m thick). f) Sediment association 2; photo shows detail of unit 4 (Section 831) shown in b. Arrows mark lag gravels at scour bases. Unit is 5 m thick. g) Erosional contact between units 3 and 4 (Section 833). h) Section 832; deposits of sediment association 1 overlain by massive diamicton. Units removed by erosion are shown in Fig. 4-5. Exposure parallels the Miette valley. Man at fabric location (Fig. 4-7b). Numbers on photos refer to unit numbers in Fig. 4-4. D, diamicton; G, gravel; S, sand.



Location Map
Pyramid Lake
Patricia Lake
Edith Lake
Annette
Athabasca River
section location
highway
road
scarp crest (approx loc)
Pyramid Thrust
Lac Beauvert
JASPER
Hwy 16
Miette R.
N
1 km
⑤ Borehole 83-5
④ Section 834
② Section 832
③ Section 833
① Section 831
UNIT 5/6
29 000 years BP
Surface Topography - actual
- before excavation
Bedrock surface - defined
- undefined
Site number (refer to Location Map)
Correlation lines
SEDIMENT TYPE
silty diamicton-matrix supported
sandy diamicton-matrix supported
gravelly diamicton-matrix to clast supported
gravels
sands
silts
covered
SEDIMENTARY STRUCTURES
planar cross-stratification
trough cross-stratification
horizontal stratification
graded bedding
channel-fill cross bedding
scour & fill structures
ripple bedding
CONTACTS
erosional
conformable
faulted
interbedded
Approximate elevation in metres
1160
1150
1140
1130
1120
1110
1100
1090
1080
1070
1060
1050
0
200
400
600
800
1000
Approximate horizontal distance in metres

Fig. 4-4 Stratigraphic columns of the main exposures in the area showing unit numbers used in text (see inset map for locations). The relative topographic positions of Sections 831 to 834 are as shown but their lateral positions are schematic (see text). Dashed lines indicate stratigraphic correlations.

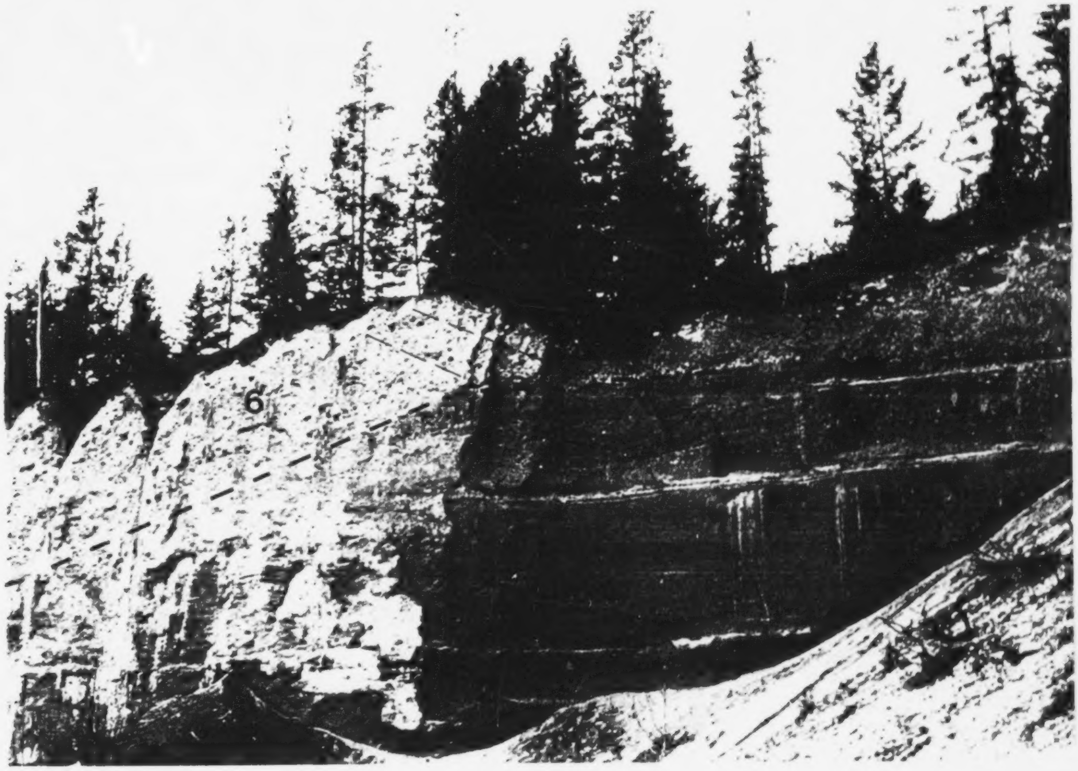


Fig. 4-5. Section 832; exposure is 19 m high and perpendicular to the valley. Massive diamicton at the top left (unit 6) grades laterally into steeply dipping (dip 30-35°, dip direction 20°), clast-supported gravels (unit 5). Note the weak stratification in the transition zone. The erosional lower contact of unit 6 dips towards the valley center (left). Exposure faces east, oblique to the Miette valley. (See also Fig. 4-3h)

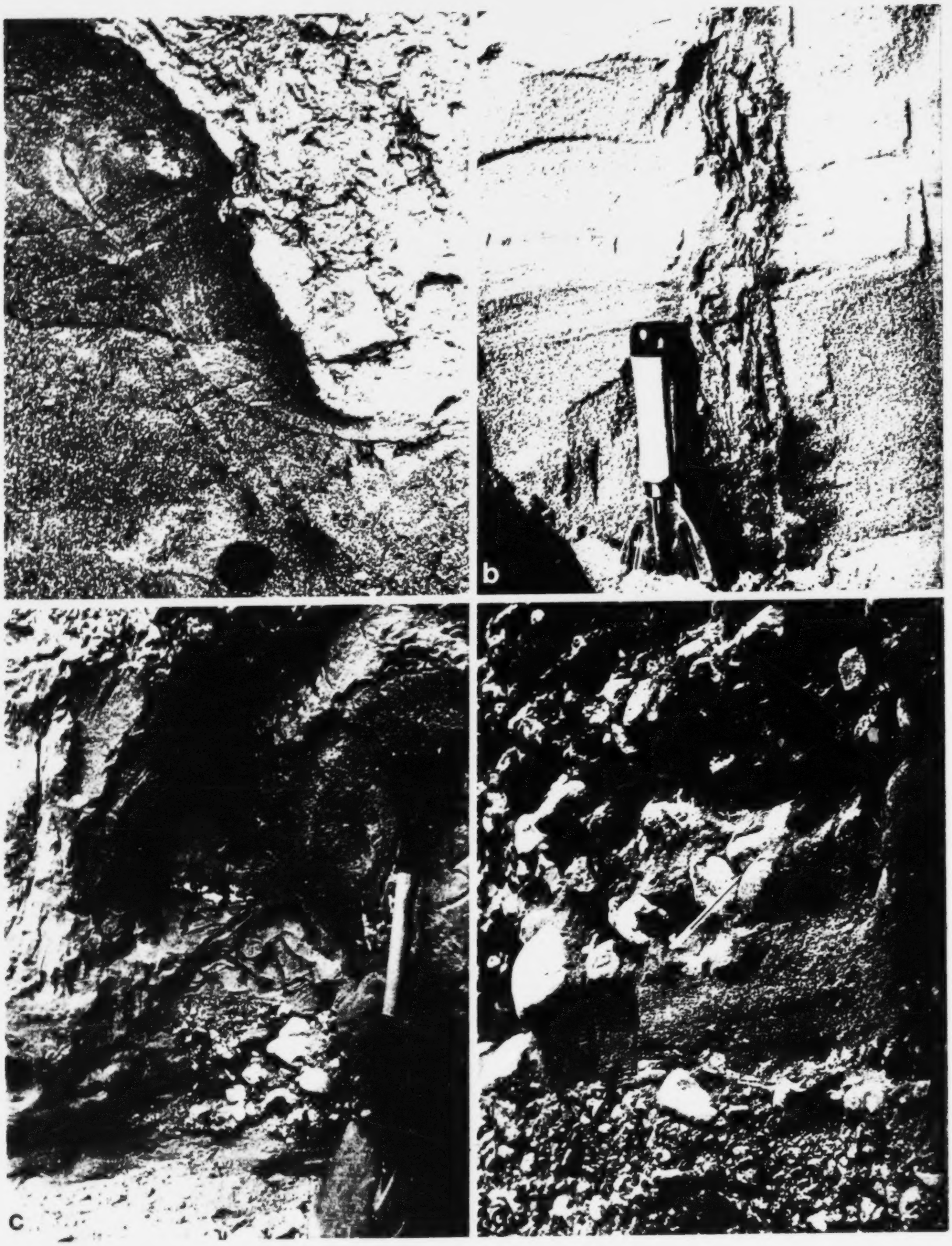


Fig. 4-6 Detailed sedimentary structures in sediment association 4. a) Deformation in sands and overlying diamicton indicating collapse towards the valley center (right), possibly due to removal of underlying ice. b) Vertically stratified silts and clays cross-cutting ripple bedded sands in unit 3, Section 834. c) Angular and deformed inclusion of gravels and sands in unit 3, Section 833. d) Contact between units 4 and 5, Section 832; horizontally stratified sands grade up into matrix-supported cobbly sands and finally into clast-supported gravels of sediment association 4.

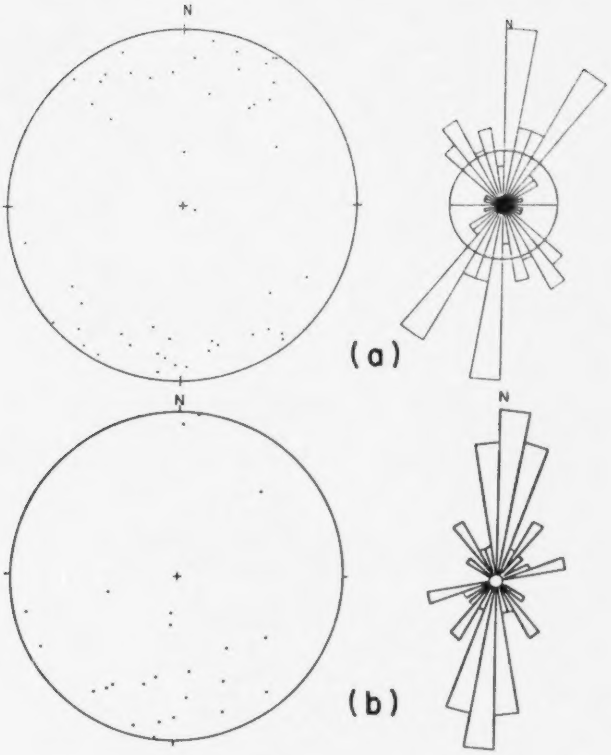


Fig. 4-7 Rose and density diagrams representing long axis orientations of medium to large, blade and prolate shaped pebbles. (V_1= orientation of maximum clustering of long axes; S_1=normalized eigenvalue for V_1; Vo= Valley orientation.) a) Diamicton of sediment association 2 (unit 4, Section 831); n=50, V_1=200°, S_1=0.65, Vo=25°. b) Diamicton of sediment association 3 (unit 6, Section 832); N=25, V_1=185°, S_1= 0.70, Vo=90.

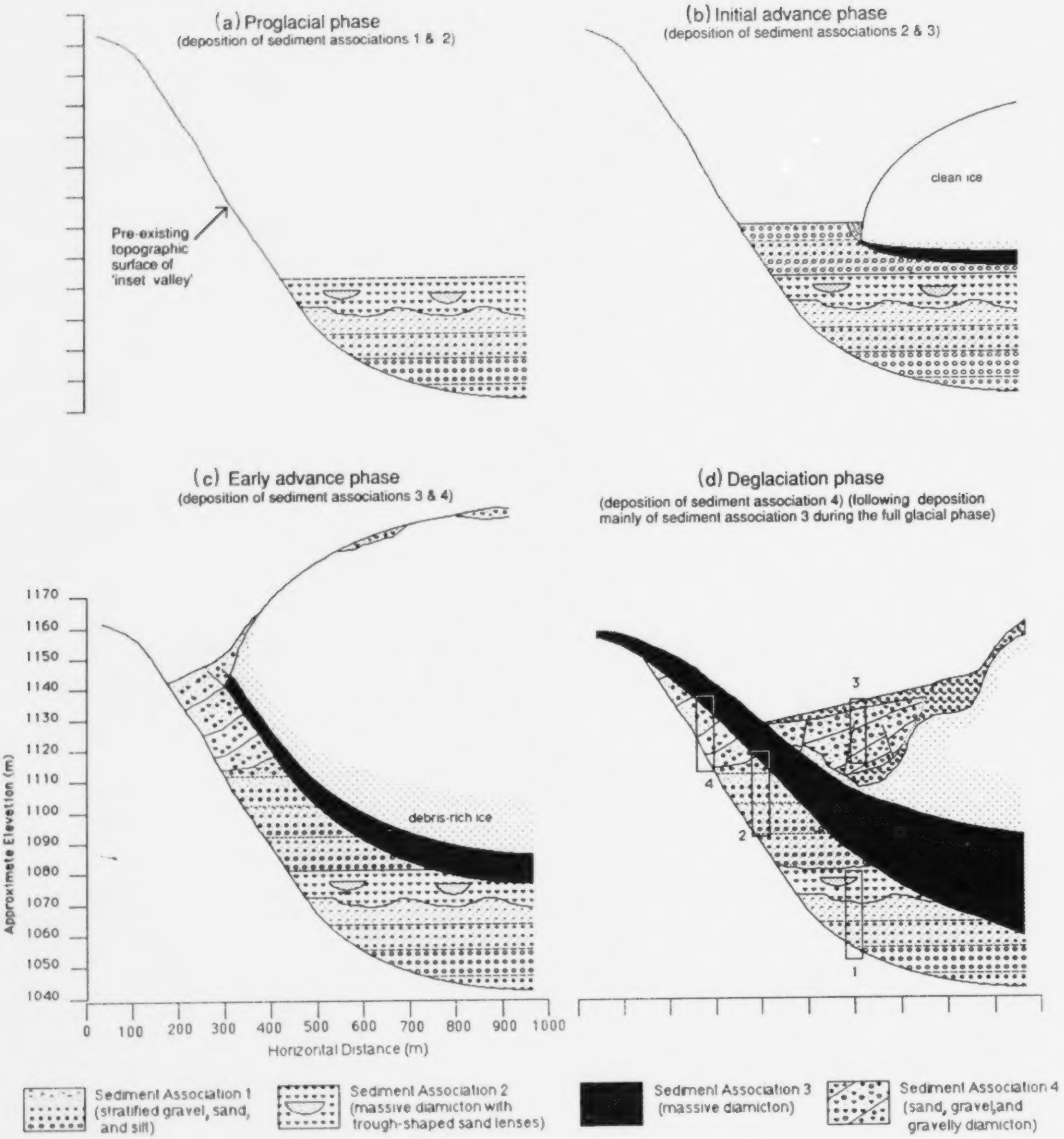

(a) Proglacial phase
(deposition of sediment associations 1 & 2)
Pre-existing topographic surface of 'inset valley'
(b) Initial advance phase
(deposition of sediment associations 2 & 3)
clean ice
(c) Early advance phase
(deposition of sediment associations 3 & 4)
debris-rich ice
(d) Deglaciation phase
(deposition of sediment association 4) (following deposition mainly of sediment association 3 during the full glacial phase)
1
2
3
4
Approximate Elevation (m)
1170
1160
1150
1140
1130
1120
1110
1100
1090
1080
1070
1060
1050
1040
0
100
200
300
400
500
600
700
800
900
1000
Horizontal Distance (m)
Sediment Association 1 (stratified gravel, sand, and silt)
Sediment Association 2 (massive diamicton with trough-shaped sand lenses)
Sediment Association 3 (massive diamicton)
Sediment Association 4 (sand, gravel, and gravelly diamicton)

Fig. 4-8 Schematic diagram illustrating the main depositional events in the formation of the terrace above Jasper townsite: a) deposition of proglacial outwash and debris flow deposits in a pre-existing inset valley, b) Athabasca Valley ice advancing down the valley center (movement direction is into the page) erodes and redeposits material at its base and confines outwash streams along the valley side, c) ice initially expands up the valley walls depositing till at the glacier base and water and gravity sorted deposits at the ice-margin, and d) glaciers retreat leaving supra- and subglacially derived tills throughout the region, and a stagnating, debris covered ice mass in the valley center from which kame terrace deposits were derived. Normal faults develop during the final meltout of the ice-shoulder. Columns numbered 1 to 4 in d) correspond to sections 831 to 834, respectively.

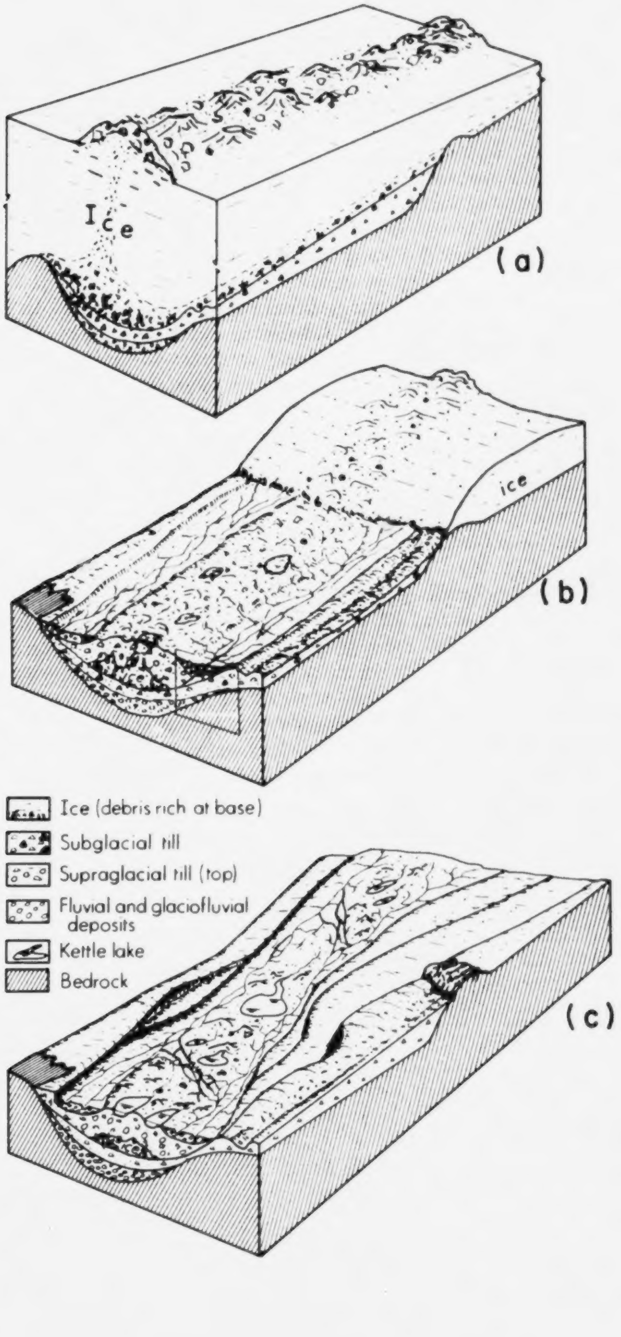


Fig. 4-9 Deglaciation of the Athabasca valley center in the vicinity of Jasper townsite. a) A well developed medial moraine forms a substantial debris cover between Athabasca (left) and Miette (right) valley glaciers. Tills deposited subglacially are thickest in the lee of bedrock highs. Note the vertical exaggeration. b) Bedrock benchlands on the valley sides become ice-free first leaving a well preserved fluted topography with little supraglacially derived material. Debris, derived from the medial moraine and by meltout of the underlying medial debris septum, accumulates on the ice surface in the valley center and eventually develops into a linear, ice-cored moraine. Ice-marginal debris flow and fluvial deposits develop into a kame terrace(s) along the sides of the decaying, debris- covered ice mass. The details of the area outlined by the box are shown in Fig. 4-8d. c) During the final melting of buried ice, a chain of kettle lakes forms in the area of the thickest supraglacial debris cover in the valley center, and anomalously large gullies, faults, and other collapse structures develop in the kame terrace deposits as underlying ice blocks decay. Note the reworking of the morainal debris by braided streams leaving a dissected and subdued kettled moraine.

Section - Unit	Deposit Type	Quartzite & Quartz Sandstone (%)	Pebbly & Granular Sandstone (%)	Siltstone (%)	Shale (%)	Limestone (%)	Dolomite (%)	Total Carbonates (%)	Schist (%)	*n*
831 - 1, 2	gravel	68	17	2	5	7	1	8	0	100
831 - 4	diamict	67	16	5	3	7	2	9	0	60
832 - 1	gravel	64	18	4	14	0	0	0	0	50
832 - 5	gravel	74	14	6	0	4	2	6	0	50
832 - 6	diamict	71	11	7	6	3	2	5	0	100
833 - 1	gravel	66	28	0	2	4	0	4	0	100
833 - 4	diamict	56	26	2	0	16	0	16	0	50
834 - 6	diamict	84	8	2	0	6	0	6	0	75
834 - 8	diamict	55	24	10	8	1	0	1	2	50

Table 4-1 Percent lithology of small to medium pebble clasts in major units in the area.

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CHAPTER 5 - EVIDENCE OF CORDILLERAN LATE WISCONSINAN GLACIERS IN THE 'ICE FREE' CORRIDOR[3]

INTRODUCTION

The extent and duration of Late Wisconsinan glaciers in the 'ice-free corridor', a belt of land between the Canadian Cordillera and the western edge of the Laurentide ice sheet, has long been discussed, particularly in relation to the importance of the corridor as a potential migration route for peopling of the New World (Rutter, 1980, 1984; Reeves, 1973, 1983). The area investigated here focuses on a critical part of the corridor in west central Alberta and adjacent parts of British Columbia where the extent and timing of ice cover is controversial (Fig. 5-1). Although most researchers in the region, particularly those working in or east of Jasper National Park (Roed, 1975; Boydell, 1978; Kvill, 1984; Levson, 1986; Levson and Rutter, 1989a, b), agree that coalescence of mountain and Laurentide glaciers occurred, the timing of this event is widely disputed. Two main schools of thought on ice position have emerged. The first favours coalescence of glaciers moving out of the Rocky Mountains with Laurentide ice in at least part of the region (*e.g.* Roed, 1975; Boydell, 1978; Mathews, 1978, 1980; Kvill, 1984; Rutter, 1980, 1984; Dyke and Prest, 1987a, b; Jackson, et al., 1989) whereas the second proposes that the corridor remained ice free for much or all of the Late Wisconsinan glacial period (*e.g.* Reeves, 1973, 1983; Stalker, 1980; Bobrowsky, 1989; Schweger, 1989; Catto and Mandryk, 1990). In order to reflect the degree of uncertainty over the extent of glaciers in the region, many researchers present both minimum and maximum ice-limit models to describe the Late Wisconsinan glacier complex (*e.g.* Prest, 1984; Fig. 5-1).

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[3] *A version of this chapter has been accepted for publication: Levson and Rutter, 1995. Quaternary International.*

Differing views on the glacial history of the region are largely due to the paucity of radiocarbon dates that adequately date the advance of ice out of the Rocky Mountains into the corridor. This problem is pronounced in the central and southern parts of Alberta where pre-Holocene radiocarbon dates on wood are lacking, and extensive Mesozoic and Tertiary coal (and other inert carbon deposits) restrict the usefulness of non-woody organics, such as peat and gytta, as sample material for radiocarbon dating (*e.g.* White *et al.,* 1985). Contamination by old carbonate or reworked organics has been discussed by numerous authors (Olsson, 1979; Mathews and Westgate, 1980; Nambudiri *et al.,* 1980; Clayton and Moran, 1982; Dyke and Prest, 1987b). This problem is further complicated when supratill radiocarbon dates from potentially contaminated lake and bog cores are used to determine the minimum date of ice retreat. Interpretations of these data have led workers to infer ice-free conditions during the Late Wisconsinan in at least three different areas (White et al., 1979; Jackson, 1980; Holloway *et al.,* 1981; Mott and Jackson, 1984). However, in each of these three cases it was subsequently demonstrated that the radiocarbon dates were too old, due to contamination of the samples by radiogenically 'dead' carbon (Hickman *et al.,* 1984; White *et al.,* 1985; MacDonald *et al.,* 1987). For this reason, emphasis is placed in this chapter on subtill radiocarbon dates on wood.

Although detailed geomorphic studies conducted in several parts of the corridor region have provided valuable information on the extent of Late Wisconsinan glaciation in the area, researchers have concluded that subtill dates from critical stratigraphic units are the best means of confirming models of ice extent (*e.g.* Rutter, 1972, 1980, 1984; Shaw, 1972; Kvill, 1984; Dyke and Prest, 1987b; Jackson, et al., 1989). The data presented here provides this type of critical evidence and supports the view that glaciers moved out of the south-central Canadian Rocky Mountains into the "ice-free" corridor during the Late Wisconsinan. In particular, the chronologic and geologic significance of a new finite radiocarbon date on wood located in sands and gravels underlying till near Jasper town site (Fig. 5-1) is discussed. This is the only finite date on wood known from east of the Rocky Mountain trench underlying Cordilleran till, that clearly indicates a Late Wisconsinan age for glaciation of the Athabasca River valley in the Jasper region. The stratigraphic and geomorphologic relationships of these glacial deposits in Jasper to the adjacent parts of the "ice-free" corridor indicate that Late Wisconsinan glaciers in the Jasper area, fed by both local Rocky Mountain and Cordilleran sources, advanced down the Athabasca valley into the corridor region where they were deflected southeasterly by Laurentide glaciers to at least the North Saskatchewan River valley area .

Geologic Setting

The southwestern part of Jasper National Park (Fig. 5-2) is located in the Rocky Mountain Main Ranges, characterized by gently dipping, thrusted and folded, Precambrian to Cambrian, quartzose sandstones, shales, siltstones and carbonate rocks. The northeast side of the park occurs in the steeply thrusted, carbonate and clastic rocks of the Front Ranges (Price and Mountjoy, 1970; Mountjoy and Price, 1985). Jasper town site is situated at the boundary between the Main and Front Ranges. The town is built on a post-glacial gravel terrace at an elevation of about 1060 metres a.s.l., approximately 30 m above the Athabasca River. Relief in the region is high, with mountain peaks commonly rising about 1500 m above valley bottoms.

Within the Main Ranges, the Athabasca River flows northwesterly along the regional structural trend. At Jasper town site, the river turns northeasterly and cuts across the strike of the Front Ranges (Fig. 5-2). The river has incised its valley through a thick fill of Quaternary glacial and glaciofluvial deposits, and is locally flanked by kame terraces and related features that rise up to about 150 m above river level (Levson and Rutter, 1989b). Northeast trending drumlins and flutings occur on the valley sides (Reimchen, 1976) and are best developed downvalley of northwest-southeasterly trending bedrock ridges.

Previous Quaternary Studies

The surficial geology of the Jasper area and regions to the east have been mapped by Reimchen (1976), Holland and Coen (1982), Roed (1970), Boydell *et al.* (1974) and Bayrock and Reimchen (1980). Roed (1975), Boydell (1978), Kvill (1984), Levson (1986) and Levson and Rutter (1986, 1989a, b; in press) have discussed the genesis of glacial deposits and aspects of the Quaternary stratigraphy of the region. Holocene glaciations in the Jasper region have been investigated by numerous authors including Luckman and Osborn (1979), Kearney and Luckman (1981), Beaudoin (1984, 1986), Luckman (1986, 1988) and Osborn and Luckman (1988).

EVIDENCE FOR LATE WISCONSINAN GLACIATION OF THE "ICE-FREE" CORRIDOR EAST OF JASPER NATIONAL PARK

Three lines of evidence suggest that Late Wisconsinan glaciers occupied the Athabasca River valley in the Jasper Park region and flowed out into the "ice-free" corridor area. These include: lithostratigraphic and chronologic (radiocarbon) data from the Jasper region that constrain the timing of the last glaciation there to the Late Wisconsinan; the distribution of metamorphic erratics in the Athabasca valley which indicate that the last glaciers to flow through the Jasper region extended well out into the ice-free corridor area; geomorphic and stratigraphic data that suggest that these same glaciers coalesced with Laurentide ice, effectively blocking the corridor region. This evidence is discussed separately in the following sections.

LATE PLEISTOCENE STRATIGRAPHY OF THE JASPER REGION

Late Pleistocene sediments in the Jasper area are exposed at a number of localities within the Athabasca River valley and its tributary valleys (Fig. 5-2). A composite stratigraphic column, summarizing the sedimentary sequence within the Athabasca valley, is shown in Figure 5-3. Five main lithostratigraphic units are identified and discussed.

Middle Wisconsinan (and older ?) Alluvial Fan Deposits

The lowermost unit in the sequence consists of poorly sorted, clast-supported, massive to crudely bedded, gravelly diamicton, interbedded with minor sands and gravels (Fig. 5-3, Plate 5-1). An alluvial fan origin for these deposits is indicated by consistently dipping strata towards the valley centre and by the dominance of clasts derived from restricted drainage areas on the adjacent valley sides. Glacial erratics are rare indicating a non-glacial origin for the bulk of the fan sediments. Crude stratification, coarse grain-size, high clast angularity and overall poor sorting of these gravelly deposits also are consistent with an alluvial fan origin. Deposition likely resulted from episodic debris and flood flows associated with the development of the fans, and to a lesser extent, from intermittent fluvial activity.

Dateable materials within the fan sediments are rare, but wood fragments (*Picea* sp.), obtained from an exposure at Amber Mountain (Site 1, Fig. 5-2), were dated at >39,750 (AECV 450c) and >48,000 (GSC-4646HP). Since the wood occurs in the upper part of the section (Plate 5-1), a large part of the fan sequence at that site must have accumulated prior to 48 ka. Deposits in the upper part of the fan sequences are locally interbedded with overlying glaciofluvial gravels indicating that alluvial fan sedimentation must have continued until the onset of the last glaciation. The age of the oldest fan deposits is unknown, but the presence of rare glacial erratics throughout the sequences indicates that they began to accumulate sometime after the first Pleistocene glaciers reached the area.

Middle to Late Wisconsinan Fluvial and Glaciofluvial Gravels

Alluvial fan sediments grade into overlying stratified gravels and sands interpreted as braided stream deposits (Fig. 5-3). Well developed pebble roundness, imbrication, stratification and sorting in these deposits and the dominance of crude horizontal bedding is consistent with deposition in a shallow, gravel-bed, braided stream environment (*c.f.* Smith, 1974; Hein and Walker, 1977). Planar cross-bedded gravels and large, trough cross-bedded, sand and gravel lenses up to several metres thick and tens of metres wide, are locally common and represent transverse bar and channel-fill deposits, respectively. These gravels are unconformably overlain by till (Plate 5-2). Organic materials, including wood and small gastropod and pelecypod shell fragments, occur throughout the lower part of this unit. Wood fragments are angular and fragile suggesting that they were not reworked from older deposits (Plate 5-3). A radiocarbon date of 29,100 ± 560 years B.P. (GSC-3792) was obtained from wood in these deposits near Jasper town site (Site 2, Fig. 5-2) at a depth of 17.9 m below the surface (1115 metres a.s.l.), indicating a Middle Wisconsinan age. The contact between the gravels and the underlying paleofan sequences is conformable as indicated by local interbedding of alluvial fan diamicton with the lower part of the gravels and a gradual vertical increase in abundance of distally derived rocks (Levson and Rutter, in press). This transition suggests that stream sedimentation was locally coeval with the latter stages of alluvial fan development.

In some areas, such as around Jasper town site and Roche Miette (Fig. 5-2), vertical facies changes in the upper part of gravels of this unit reflect the increasing proximity of glaciers. These changes include a vertical decrease in the abundance of wood and shell fragments and an upward increase in the number of cut-and-fill structures, erratic clasts, coarsening-up beds, poorly sorted gravels and gravelly diamictons. These features are typical of gravelly, ice-proximal, braided rivers (Rust, 1972, 1978; Boothroyd and Ashley, 1975; Church and Gilbert, 1975). In some areas, diamicton beds, containing striated clasts of distal lithology, occur in the uppermost part of this unit. The diamicton beds are interpreted as debris flow sediments deposited in the ice-marginal streams of advancing glaciers. Interbedded gravel and sand strata show large variations in sediment sorting, grain size distribution, maximum clast size, clast shape and paleocurrent direction. These variations indicate proximity to a sediment source and highly variable flows, typical of ice-marginal environments (Boulton and Eyles, 1979). The gravels locally grade towards the valley center into a massive diamicton unit that has been interpreted as till (Levson and Rutter, 1989b). Bedding in the upper few metres of the gravels is locally deformed, with overturned folds and thrust faults. Deformation is most intense in the vicinity of large 'embedded' boulders with faceted and striated upper surfaces. This deformation also suggests that the gravels were overridden by glaciers.

Late Wisconsinan, Advance-Phase Glaciolacustrine Sediments

In the lower reaches of several tributary valleys of the Athabasca River such as the Jacques, Snake Indian and Rocky River valleys (Fig. 5-2), glaciofluvial gravels are overlain by horizontally stratified silts, clays and fine sands (Fig. 5-3). These sediments are rhythmically bedded, contain numerous dropstones and commonly coarsen upwards. They are interpreted as glaciolacustrine deposits and are invariably overlain by glacially derived debris flow deposits and tills (see below). Deposition presumably occurred in lakes formed in tributary valleys where drainage was dammed by glaciers in adjacent trunk valleys. Ice-dammed lakes were a characteristic feature of tributary valleys in the study area at the onset of the last glacial period, particularly in Front Range valleys that remained ice free during the early phases of glaciation (Levson, 1990).

Late Wisconsinan Glacial Sediments

Basal Tills

Thick sequences of diamicton, interpreted as basal tills, occur at or near the surface throughout the Jasper area (Plate 5-1). They unconformably overlie bedrock or the Quaternary sequences discussed above. The diamictons are mainly fine-grained, massive, matrix-supported, unsorted and over-consolidated. The tills commonly show fluted or drumlinized upper surfaces, strong downvalley pebble fabrics, subhorizontal shear structures, sharp erosional lower contacts and numerous lodged and striated clasts. These characteristics are typical of basal tills deposited by both lodgment and meltout processes (Boulton, 1976; Haldorsen, 1982; Haldorsen and Shaw, 1982; Shaw, 1982; Dreimanis, 1989). Sedimentary structures suggestive of lodgment processes include numerous faceted clasts that are embedded (lodged) in compressively deformed underlying sediments, locally common slickensided shear planes and abundant heavily striated clasts, with striation and long-axis trends generally parallel to ice-flow direction (Levson and Rutter, 1989a). Inferred meltout structures include rare circular (englacial) or plano-convex (subglacial) sand and gravel lenses, often with diamicton inclusions or with large clasts that protrude into the lenses from the encasing diamicton, and diffuse diamicton strata (possible relict ice-debris bands) that uniformly drape underlying boulders. The diamictons were deposited mainly by Athabasca Valley ice but the influence of tributary-valley glaciers is locally indicated by stratigraphically complex diamicton sequences of variable provenance (Levson and Rutter, 1986). The presence of fluted and drumlinized upper surfaces clearly indicates an overriding glacier, with the streamlined forms reflecting subglacial water and/or ice erosion.

Supraglacial Tills

Sandy, poorly consolidated, diamicton, containing angular to subrounded clasts up to boulder size, locally occurs on the surface in the Jasper area and often overlies massive diamictons interpreted as basal or englacial meltout tills (Levson and Rutter, 1989a). Diamicton beds are often interbedded with sands and gravels and are usually poorly exposed due to the loose sandy texture and relative thinness of the deposits (generally up to about one meter thick). The diamictons occur sporadically and are represented by scattered metamorphic erratics that may directly overlie bedrock. These erratics rarely occur in subsurface diamictons. They are interpreted to be part of the Athabasca Valley erratics train described by Roed *et al.* (1967) from the area east of Jasper (see discussion below). Analogues for these isolated supraglacial erratics can be found on modern valley glaciers in the region (Plate 5-4c).

The sandy matrices, poor consolidation, high clast angularity and large modal clast size of the diamictons, and their common vertical association with underlying meltout till deposits, suggests that they were supraglacially derived. Similar characteristics have been described for supraglacial tills by numerous authors (Sharp, 1949; Drake, 1971; Stewart and MacClintock, 1971; Boulton and Eyles, 1979; Lawson, 1979a, 1981a and 1981b). The presence of distally derived (>50 km) metamorphic erratics, the relative thinness of the diamictons, and their common occurrence on the surface further supports a supraglacial origin. Deposition most likely occurred in the form of debris flows off the glacier surface or through gradual let-down by melting during deglaciation. Interbedded sand and gravel deposits were probably deposited by meltwaters between debris flow events.

Ice-Marginal Deposits

Diamicton beds, with crude stratification, relatively high clast content, a sandy matrix, and poorly developed fabric, commonly cap more compact, massive, diamicton beds. In addition, they often contain interbedded sand and gravel lenses that display variable dips and dip directions. These characteristics are common in debris flow deposits in glacial environments (Marcussen 1975; Lawson, 1979a, b, 1981a, b, 1982, 1989; De Jong and Rappol, 1983). Interstratified lenses and beds of gravels, sands and silts often dip steeply (up to 35°) towards the valley sides or upvalley. Normal faults and deformation structures indicating collapse towards the valley centre occur in these deposits as do load and injection structures. These sediments are common in terrace-like benches flanking the Athabasca valley. They are interpreted as ice-marginal glaciofluvial and debris flow sediments deposited along the lateral margins of debris covered ice situated in the valley centre. Boulton and Eyles (1979) and Bordonau (1993) described similar ice-marginal deposits along the flanks of former valley glaciers. The debris source for these features is attributed to ablating medial moraines and medial debris septa that extended into the Athabasca valley from its tributaries (Levson and Rutter, 1989b). This interpretation is supported by the occurrence of terraces below major stream confluences such as the Miette, Maligne, Snaring, Rocky and Snake Indian Rivers where medial moraines and debris septa would have been best developed. Kame terraces throughout the Athabasca valley in Jasper Park and east to the Hinton area (Fig. 5-2) are compositionally and sedimentologically similar and most show similar degrees of post-glacial, geomorphic modification.

Late Wisconsinan Retreat-Phase Glaciolacustrine Sediments

At a few sites, glacial deposits overlain by massive to horizontally laminated silts and clays (Fig. 5-3; Plate 5-5), are interpreted as glaciolacustrine sediments deposited in ice-marginal lakes during retreat of the glacier occupying the Athabasca Valley. In the Pocahontas area (Site 3, Fig. 5-2), the silts and clays occur near the top of an incised bench on the southeast side of the valley. These ice-marginal sediments contain numerous gastropods of a variety of species dated at 11,900 ± 120 years B.P. (GSC - 3885), implying a maximum date for retreat of ice from the region (Levson *et. al.,* 1989).

Summary of Late Pleistocene History of the Jasper Region

Lithostratigraphic and chronologic data from Quaternary sediments in Jasper National Park indicate that ice-free conditions persisted in the region prior to 48 ka and continued beyond 29 ka, during which time extensive alluvial fans developed along the valley margins and braided streams occupied the valley bottoms. The onset of the last glaciation is recorded by glaciofluvial and glaciolacustrine sedimentation in most valleys and later by the deposition of glacial deposits throughout the region. These deposits include both basal tills and an overlying supraglacial till derived from west of the Continental Divide. Two ^{14}C dates effectively bracket this glaciation. A date of 29,100 ± 560 years B.P. (GSC 3792) obtained on wood from gravels underlying till near Jasper town site, suggests that ice moved into the region after that time. Gastropods from ice-marginal lake sediments in the Pocahontas area (Fig. 5-1), dated at 11,900 ± 120 years B.P. (GSC 3885), provide a minimum date for deglaciation. Late Wisconsinan glaciers in the Jasper region had receded to near their present limits by about 10 ka (Luckman and Osborn, 1979; Osborn and Luckman, 1988). Only minor advances, extending about one to two kilometers beyond present glacier margins, have since occurred (Kearney and

Luckman, 1981; Osborn and Luckman, 1988). During deglaciation, ice stagnated along several parts of the Athabasca valley and ice-marginal kame terraces formed along the valley walls. Terminal moraines, lateral moraines or other features indicative of the terminus of a major glacial advance of Cordilleran ice are not observed (Levson and Rutter, in press). These data lead to the conclusion that Cordilleran glaciers moved down the Athabasca valley during the Late Wisconsinan and terminated somewhere east of the Front Ranges in the "ice-free" corridor region.

ATHABASCA VALLEY ERRATICS

Throughout the Athabasca River valley downstream of the Miette River confluence, angular metamorphic erratics (mainly talcose muscovite and garnet schists with lesser amounts of biotite schist and hornblende-garnet-quartz gneiss) locally occur on till surfaces and on surrounding bedrock ridges up to an elevation of about 2000 metres. In the Main Ranges near Jasper, muscovite-garnet schist and quartz-biotite gneiss erratics occur up to 2240 metres a.s.l.. Talcose mica and garnet schists are found up to 1600 metres in the Front Ranges and Foothills. The erratics occur in association with supraglacially derived, debris-flow deposits. Many of the associated metamorphic erratics are soft and friable due to their well developed foliation and they are unlikely to have survived transport for any significant distance except in a supraglacial or englacial position. The rarity of these erratics in till deposits below the surface supports this interpretation.

Similar metamorphic rocks, described from the region east of Hinton, are referred to as the Athabasca Valley Erratics Train (Roed *et al.*, 1967). They occur at elevations of 1000 to 1250 metres and extend, in the area north of the Athabasca River, approximately 50 kilometres beyond the edge of the Rocky Mountain Foothills and as far south as latitude 53° N (Roed, 1975). Garnet-talcose schists, ascribed to the Athabasca Valley

Erratics Train, also occur along the mountain front in the Brazeau and North Saskatchewan River areas (Boydell, 1978). Roed *et al.* (1967) concluded that these metamorphic erratics were carried by the last glaciers that flowed across the Continental Divide into the Miette and Athabasca valleys (Fig. 5-1). The glaciated 'U-shape' of the main passes along the Continental Divide west of Jasper such as Yellowhead Pass (Fig. 5-1) attest to this movement of ice. The erratics have probable source rocks in British Columbia west of the Rocky Mountain Trench in the Monashee Mountains or Premier Range or possibly along the east side of the trench where the garnet and biotite isograds cross the west slopes of the Selwyn Mountains (Price *et al.,* 1977). Hornblende gneiss erratics, in particular, indicate a source at the northwest end of the Monashee Mountains in the Malton Range (Roed *et al.,* 1967).

The occurrence of the Athabasca Valley Erratics Train within Jasper Park and east of the Rockies, indicates that glaciers flowing down the Athabasca River valley extended well out into the "ice-free" corridor region. The possibility that the train was deposited during more than one Cordilleran glaciation, by ice that moved across the Continental Divide after having eroded and transported an identical suite of metamorphic erratics, seems unlikely. The erratics everywhere occur at the surface, do not show noticeable differences in weathering across the region and are associated with correlative till units. The stratigraphic position, angularity and soft, non-resistant nature of the erratics precludes erosion from an older till and redeposition at the surface by a separate overriding glacier. Therefore, transportation and deposition of the erratics is presumed to have occurred during the latest glacial event, at a time when Cordilleran glaciers flowed down the Athabasca valley, out of the Rocky Mountains and then southeastward, at least as far as the North Saskatchewan River valley. This is also supported by the continuity and geomorphic similarity of the surficial tills associated with the erratics, as discussed below.

GLACIAL DEPOSITS EAST OF JASPER PARK

The nature and distribution of glacial deposits of Cordilleran provenance, along the east flank of the Rocky Mountain Foothills, provides further evidence that glaciers extended well east of the mountain front during the last glaciation. These deposits can be traced from the Jasper region for 200 km or more. They are characterized by a remarkably concordant pattern of flutings and drumlins that trend northeasterly along the Athabasca valley in the Hinton area and then gradually shift to the east and southeast, paralleling the Rocky Mountain Foothills (Fig. 5-4). Disruptions in this general pattern are almost entirely restricted to areas where major rivers flow from the Rocky Mountains. Throughout this region, the Athabasca Valley till is lithologically similar and contains erratics that indicate a Cordilleran provenance.

East of the Rocky Mountain front in the Edson-Hinton area, three glacial tills derived from the mountains, termed the Marlboro, Raven Creek and Obed tills are recognized (Roed, 1970, 1975). All three tills are considered to be Late Wisconsinan in age. The Marlboro and Raven Creek tills represent lateral facies equivalents deposited during a single glacial event, whereas the Obed till was deposited during a readvance (Roed, 1975). The Marlboro facies is widespread, occurring as a broad lobate deposit northeast, east and southeast of Hinton. Clasts within the till are mainly carbonate and clastic rocks derived from the Rocky Mountains, with minor metamorphic rocks of the Athabasca Valley Erratics Train. The Marlboro till facies grades laterally to the southeast into the Raven Creek facies which is distinguished only by the presence of rare (up to 2%) Canadian Shield erratics. The two tills are interpreted by Roed (1975) as deposits of the same glacier on the basis of their compositional similarity, large variations in elevation (up to 300 metres) along their lateral contact which suggests a single ice mass, and the lack of any discordance in the trends of subglacial paleoflow indicators between them.

The Obed till is lithologically similar to the Marlboro till but is locally more sandy and contains a higher proportion of limestone pebbles. The surface expressions of the two deposits are similar but they can be differentiated morphologically in places. In the Hinton area, for example, northeast-trending drumlins and flutings in Obed deposits along the Athabasca valley appear to truncate southeast-trending features in the Marlboro till to the south (Fig. 5-4). Based on these observations and stratigraphic correlations, Roed (1975, p. 1512) suggested that the Obed till was likely deposited during a glacial readvance at the end of the Marlboro phase, and incorporated ice left from the Marlboro glacier. This interpretation is supported by the occurrence of talcose and garnet schists of the Athabasca Valley Erratics Train on the surface of both tills. It is unlikely that the incompetent schistose erratics could have survived erosion from the Marlboro till, been transported and later redeposited at the surface by a subsequent glacial event (Roed, 1975). The evidence presented by Roed (1970, 1975), suggests that the Obed, Marlboro and Raven Creek units were all deposited during the same glacial period by Cordilleran ice flowing down the Athabasca valley. Kvill (1984) reached the same conclusion after conducting geomorphic, aerial reconnaissance, airphoto and surface investigations in the Athabasca valley region. Although supportive evidence has not been found by other researchers (*e.g.* Kvill, 1984), Roed (1975) proposed a subsequent minor advance, possibly occurring during retreat of Obed ice, that was considered to be short-lived and restricted to local alpine glaciers in the Front Ranges that did not extend into the main valleys.

The glacial geomorphology of the Brazeau River area (Fig. 5-4) was investigated by Kvill (1984). Mainly on the basis of geomorphic data, glacial deposits in the region were attributed to a single glacial event. Geomorphic evidence supporting a single event includes a similar degree of landform modification by non-glacial processes throughout the area. For example, the relative development of drainage networks and the degree to which the topography has been subdued by colluviation, does not vary throughout the region as would be expected if glaciers had terminated somewhere in the study area. In addition, soil profile development and weathering depths are similar throughout the area. End moraines and related geomorphic boundaries are lacking except within a few kilometres of present icefields (Kvill, 1984).

The surficial geology of the area south of latitude 53° and east of longitude 116° (Fig. 5-4) was mapped by Boydell *et al.* (1974). Till deposited by glaciers flowing down the North Saskatchewan River valley is the dominant glacial deposit in the west part of the region, containing mainly carbonates and quartzites from the Rocky Mountains and locally derived, Mesozoic sandstones. This till can be traced from the lower North Saskatchewan River valley into the Rocky Mountains and, also on the basis of stratigraphic equivalence, surface tills in the two areas are considered to be correlative (Boydell, 1978). Older tills in both areas are recognized only in the subsurface record (McPherson, 1970; Boydell, 1978). In the Rockies, the surface till overlies thick deposits of glaciofluvial sand and gravel (Plate 5-6), forming a stratigraphic sequence similar to that in the Athabasca River valley. Glacial deposits east of the Rockies derived from the North Saskatchewan valley are distinguished from Athabasca valley till mainly by their lithologic and heavy mineral suites (Boydell, 1978). The Athabasca valley till in this area contains carbonate and clastic pebbles from the Rocky Mountains, and minor garnet-talcose schists ascribed to the Athabasca Valley Erratics Train. The till is inferred to be a southeastward facies equivalent of the Raven Creek and Marlboro units, all three units being deposited by the same Cordilleran glacier (Roed, 1975; Boydell, 1978).

DISCUSSION

EVIDENCE FOR COALESCENCE OF ATHABASCA RIVER VALLEY GLACIERS AND THE LAURENTIDE ICE SHEET

Evidence that the Athabasca valley till and Laurentide tills to the east were deposited by confluent glaciers, is given by morphologic, stratigraphic and lithologic data. Geomorphic evidence for coalescence is provided mainly by regional patterns of subglacial paleoflow indicators (Fig. 5-4). In the Athabasca River valley, Roed (1975) documented coalescence on the basis of fluting and drumlin patterns that suggest deflection of mountain (Marlboro-Obed) glaciers by the Laurentide ice sheet. The regional pattern of fluting and drumlin orientations in the Athabasca valley till south of the Athabasca valley shows a well developed southeasterly trend (Fig. 5-4). If glaciers were allowed to flow unconfined from the mountains down the regional topographic slope, a northeasterly orientation would be predicted. The southeasterly regional trend is not disturbed by major cross-cutting valleys, such as the Brazeau River valley. Fluting orientations in that area do not shift from southeasterly to easterly, as would be expected if down-valley deflection occurred, but instead they shift to the south, upslope towards the Foothills (Fig. 5-4). In addition, the orientations of drumlins and linear ridges in the North Saskatchewan River valley, shift gradually from northeasterly in the mountains, to easterly, southeasterly and eventually southerly at progressively lower elevations in the North Saskatchewan River valley (Fig. 5-4). This change was attributed to an ice-flow deflection caused by glaciers further east that deposited the Athabasca valley till (Boydell *et al.,* 1974; Boydell, 1978).

Although drumlins and flutings are commonly considered to be glacial erosional or depositional features, their formation by catastrophic, subglacial, water flows has been proposed for several areas of Laurentide glaciation (*e.g.* Shaw, 1983, 1989; Shaw and

Kvill, 1984; Shaw and Sharpe, 1987; Shaw *et al.*, 1989; Shoemaker, 1992a, b). Rains *et al.* (1993) argued that drumlin and fluting fields in the present study area were also formed by subglacial waters and they also concluded that these features formed under Late Wisconsinan piedmont glaciers flowing out of the mountains that coalesced with the Laurentide ice sheet to the east. Regardless of whether these features formed by subglacial ice or water erosion or by a combination of both processes, the evidence presented here suggests that they developed in a zone of coalescence of Laurentide and mountain glaciers.

Lithologic evidence for coalescence is provided by the distribution of Cordilleran tills in the region which not only indicates that Cordilleran glaciers flowed out into the "ice-free" corridor, but also suggests that these glaciers were confined by Laurentide ice to relatively high elevations along the margins of the Rocky Mountain Foothills, instead of flowing to the east as dictated by the regional topography. The distribution of metamorphic rocks of the Athabasca Valley Erratics Train, southeastward along the mountain front rather than eastward along the Athabasca valley, also indicates that Cordilleran glaciers were deflected southeastward by the Laurentide ice sheet.

Glacial flow indicators, stratigraphic data, geomorphic projections and regional similarities in till character, were used by Roed (1975) and Boydell (1978) to show that the Athabasca River and North Saskatchewan River areas were connected during the Late Wisconsinan by a southeasterly flowing, continuous ice-mass that was confluent with Laurentide ice to the east. Based on geomorphic evidence obtained from the intervening region, including the presence of glaciolacustrine sediments perched on the eastern side of the Foothills, the orientations of ice-moulded landforms, and the locations of meltwater channels and moraines, Kvill (1984) concluded that glaciers flowing down the Brazeau River valley were confluent with glaciers diverted southward along the foothills from the Athabasca valley and the adjoining Laurentide ice sheet to the east.

Changes in the lithologic composition of the Athabasca Valley till and the nature of its lateral contact with Laurentide till, indicate that the associated ice masses were confluent. For example, the proportion of Canadian Shield rocks in the Athabasca valley till progressively increases to the southeast, from 0% along the Athabasca valley itself, to 2% at 53° N latitude, to 10% along the North Saskatchewan River valley and finally to 15% at 52° N latitude. This probably reflects progressive incorporation of shield erratics from older deposits as the glacier system flowed southeastward. In the area where Cordilleran glaciers would have first merged with the Laurentide ice sheets (between latitudes 53° 30' N and 53° N, Fig. 5-4), the contact between Athabasca valley till and Laurentide glacial deposits to the east is sharp (Roed, 1975). Further to the southeast, this contact becomes more diffuse but on either side of the contact zone, the two tills can be clearly differentiated by pebble lithology data. South of 53° N latitude, evidence that Athabasca and Laurentide tills were deposited by confluent glaciers is provided by the stratigraphic and morphologic continuity of the two tills, a gradational change in composition with no apparent break in lithologic characteristics, a lateral contact not marked by any distinctive feature and the observation that the tills are not superimposed or overlapping (Boydell, 1978).

The possibility that Late Wisconsinan glaciers in the Brazeau River region did not coalesce with the Cordilleran glaciers from the Athabasca valley (either because they were restricted to the Rockies and did not extend far enough into the Foothills or because maximum extension of the two ice masses was non-synchronous) was negated on the grounds of several arguments (Kvill, 1984). These include the absence of end moraines marking the downvalley limit of Brazeau River valley glaciers, and similar non-glacial topographic modification and drainage network development. In addition, soil profile development and weathering depths are essentially the same in valleys both within and east of the Rockies. Glacial landforms such as moraines and eskers in the Brazeau River valley in the Foothills are not draped by glacial lake sediments as would be expected if the valley was ice-free when the Athabasca lobe blocked drainage to the east. Finally, the distribution of moraines and other landforms in the Brazeau valley do not suggest that Athabasca lobe ice flowed up into the valley as would be expected if the valley was ice free.

Evidence for coalescence of glaciers flowing out of the Rocky Mountains with Laurentide glaciers is also provided by large supraglacial erratics of the Foothills Erratics Train (Plate 5-4b) which occur east of the Rocky Mountain Foothills from the Athabasca River to south of the Canada-U.S. international border (Stalker, 1956; Mountjoy, 1958; Klassen, 1989). Since these quartzitic erratics are derived from the Rockies and occur along a southeasterly trending belt, rather than being restricted to mountain valleys, they were likely transported by mountain glaciers confined to higher elevations to the west by the Laurentide ice sheet. In addition, the Foothills Erratics Train was not overridden by either Laurentide or mountain ice masses, and therefore, must have been deposited during the last glacial event, with retreat of the two ice masses occurring at about the same time (Clague, 1989). The association of 'mixed provenance' tills containing shield and mountain erratics with the Foothills Erratics Train also provides evidence that the train was deposited by mountain glaciers confluent with Laurentide ice (Jackson *et al.*, 1989).

Further evidence for Cordilleran and Laurentide glacier coalescence is indicated elsewhere in the "ice-free" corridor on the basis of geomorphic and stratigraphic data (Rutter, 1984). In the Peace River valley, north of the area investigated here, coalescence of the two ice masses during the Late Wisconsinan has been suggested for more than a decade (Rutter, 1976, 1977, 1984; Mathews, 1978, 1980), although recently questioned by Bobrowsky (1989). The area of proposed confluence is marked by a continuous till sheet with granite and gneiss shield erratics in the east, and Cordilleran (schist or slate) and Rocky Mountain erratics in the west (Mathews, 1978). The till sheet occurs at approximately the same elevation and is not marked by features such as end moraines that would indicate non-coalescence (Rutter, 1984). Fluting patterns indicate that

Laurentide ice, flowing southwest into the Peace River area, was deflected south and later southeast by ice flowing from the Rockies (Mathews 1980) .

TIMING OF GLACIATION IN JASPER AND ADJACENT AREAS

The Quaternary stratigraphy of the Jasper region and radiocarbon chronology of the area, based on dates from wood underlying till, clearly indicates that ice-free conditions persisted in the Athabasca Valley throughout most of the Middle Wisconsinan, from at least 48 ka to sometime after 29 ka. Sub-till wood dates, as young as 19,100±850 B.P. (GX-2033) in the southern Rocky Mountain Trench (Clague, 1975) and 15,180±100 B.P. (TO-708) in the Finlay River area in northern British Columbia (Bobrowsky and Rutter, 1992), indicate that Cordilleran ice probably crossed the trench in those areas in the latter part of the Late Wisconsinan. Four dates on wood in the trench near Mica Dam, about 75 km west of the Columbia Icefield (Fig. 5-2), range from 21,500±300 B.P. (GSC-173) to 25,200±260 B.P. (GSC-1802), suggesting that ice-free conditions persisted there during that period (Fulton and Archard, 1985). Cordilleran glaciers from west of the Rocky Mountain Trench, therefore, probably did not move into the Jasper area before 21.5 ka. However, glaciers from nearby sources in the Rocky Mountains, and possibly the Selwyn Mountains in British Columbia, most likely flowed down the Athabasca Valley prior to this time. This is implied by nearby icefield accumulation areas, currently supporting active glaciers, that occur within the upper reaches of the Athabasca River valley and its tributary valleys as near as 20 kilometres from Jasper town site. Pebble fabric data, lithologic studies, and stratigraphic and geomorphic evidence, indicate that Athabasca valley glaciers moved into the Jasper town site area prior to glaciers from the west (Levson and Rutter, 1989b).

The radiocarbon chronology of the Cordilleran and Laurentide ice sheets in other regions, as defined by subtill wood dates, also agrees well with the conclusion that

Cordilleran glaciers moved across the Rockies during the Late Wisconsinan. Chronologic control on the coalescence of Cordilleran and Laurentide glaciers in the Peace River area in northeastern B.C. was provided by a date of 27,400±580 B.P. (GSC-2034) on a tooth from gravels predating the advance of ice into Fort St. John area (Mathews, 1978). A radiocarbon date of 25,940 ± 380 (GSC-573) on plant matter underlying till in the Finlay River valley west of Fort St. John was used for many years to date this event (Rutter, 1976, 1977, 1980, 1984). This was recently corroborated by a series of radiocarbon dates, ranging from about 15 ka to 37 ka, on wood and peat underlying till in the same valley (Bobrowsky, 1989; Bobrowsky and Rutter, 1992). The youngest date was used to argue for a short-lived Late Wisconsinan glaciation of limited extent. The timing of the Late Wisconsinan, Laurentide advance east of the corridor has been well constrained by radiocarbon dates (Westgate *et al.*, 1972; Liverman *et al.*, 1989; Liverman, 1990). Six finite dates from wood and two from organic detritus, give an age range of 27 ka to 43 ka for the sediments that stratigraphically underlie till at Watino (Fig. 5-1). Evidence presented by Liverman *et al.* (1989) suggests that the Late Wisconsinan represents the only Laurentide glaciation of the Watino area. Young *et al.* (1994) presented similar evidence for the Edmonton area. If these conclusions are correct, and barring the unlikely possibility of an extremely lobate western margin to the Laurentide ice sheet (Fig. 5-1), then confluence of Laurentide and Cordilleran glaciers in western Alberta must have occurred in the Late Wisconsinan.

A Late Wisconsinan age for glacial deposits in the Brazeau River region is suggested by a radiocarbon date of 11 225±125 years B.P. (S-1705) on organic sediments just above the transition between glaciolacustrine and lacustrine deposits in Fairfax Lake, located north of the Brazeau River near the mountain front (Kvill, 1984). The relatively low amount of pollen and the plant assemblage represented in the lower part of the core were interpreted to indicate the establishment of a pioneer plant community immediately following deglaciation. On the basis of these observations, Kvill (1984) concluded that the glacial features in the Brazeau River region were a result of glaciation during the Late Wisconsinan. The pollen record from Fairfax Lake is similar to that of Goldeye Lake, located near the mountain front about 10 km southwest of Nordegg (Fig. 5-4), where the basal pollen zone is dated from >24 to 11.4 ka (Schweger, 1989). However, correlation with the nearby Nordegg Pond pollen record, where much younger basal dates on terrestrial macrofossils have been recovered (Mandryk, 1992), suggests that radiocarbon dates on the lower pollen zone at Goldeye Lake may be contaminated.

In the North Saskatchewan River valley, Boydell (1978) presented evidence, including the stratigraphy of tills and associated glaciolacustrine deposits, and geomorphic data, to demonstrate the synchronous presence of mountain glaciers and the Laurentide ice sheet during the most recent glacial event in the area. A Late Wisconsinan age was assigned to this event. No sub-till radiocarbon dates were available, but dates of 9670±140 BP on bison bone and 10,250±165 BP on gastropod shells, obtained from lake sediments in a stagnant-ice complex, were used to corroborate a Late Wisconsinan age for this event (Boydell, 1978, P. 29). Mandryk (1990) presented evidence for a much earlier deglaciation at Mitchell Lake, where *Sphaerium* and *Pisidium* shells, obtained from a lake sediment core, yielded accelerator dates of 17,960±160 B.P. (TO-574) and 14,740±130 B.P. (TO-573). The lake occurs immediately south of the study area, near the boundary of the Athabasca and North Saskatchewan Valley tills, where early deglaciation is expected. Paleoecological data indicate a cold, semiarid climate with low primary productivity until about 11 ka (Mandryk, 1992). Although the possibility of contamination of the older dates at Mitchell Lake was discussed and considered unlikely by Mandryk (1990), it has not been entirely ruled out (Mandryk, personal communication, 1994).

Other than chronologic data from the areas discussed above, radiocarbon dates that constrain the timing of coalescence of Cordilleran and Laurentide glaciers in the corridor region are lacking. Roed (1975), for example, assigned the Marlboro-Obed glacial to the Late Wisconsinan on the basis of regional correlations but no organic materials were discovered under the associated tills for radiocarbon verification. On the basis of geomorphic, stratigraphic and other data, most workers have concluded that coalescence occurred during the Late Wisconsinan. Rutter (1980, 1984), for example, summarized previous studies throughout the corridor region and concluded that the available evidence favoured ice occlusion during the Late Wisconsinan, at least in parts of west-central Alberta and northeast British Columbia. The only data presented that questioned this interpretation were two radiocarbon dates of 17, 570±650 B.P. (WAT-406) and >30,000 BP (WAT-361) on organic sediments from a core at Boone Lake in the corridor region in northwestern Alberta, 110 km southeast of Fort St. John (White *et al.*, 1979). These dates were later demonstrated to be spuriously old due to contamination by Cretaceous organic matter (White *et al.*, 1985).

Similarly, Klassen (1989) summarized the Quaternary geology of the southern corridor region and concluded that an extensive Late Wisconsinan ice cover (see maximum portrayal limit on Fig. 5-2) was more probable than a restricted cover. This was based on a radiocarbon date of 12 630±80 BP (TO-216) on freshwater shells from deposits in a former ice-walled lake located 75 kilometres beyond the minimum ice limit. Basal radiocarbon dates from a peat bog outside the minimum ice limit in southwest Alberta, originally led researchers to conclude that the area was ice-free about 18 ka (Jackson, 1980; Mott and Jackson, 1982), but these dates were also later demonstrated to be several thousand years too old (MacDonald *et al.*, 1987). In a review of this information, Jackson *et al.* (1989) concluded that glaciers flowing out of the Rockies during the Late Wisconsinan coalesced with Laurentide glaciers in the region from at least the Athabasca River in the north to beyond the Bow River in the south.

EXTENT OF LATE WISCONSINAN GLACIATION IN THE CORRIDOR REGION EAST OF JASPER

The extent of Late Wisconsinan ice cover in the Rocky Mountain Foothills and Front Ranges, east of Jasper Park between the Athabasca, Brazeau and North Saskatchewan River valleys is less clear than in the corridor region east of the Foothills. Biological evidence indicates that some high elevation areas in this region were ice-free during the Wisconsinan (Packer and Vitt, 1974; Clifford and Bergstrom, 1976; Packer, 1980; Mayhood, 1991) but the extent of these refugia is not well documented. A steep aridity gradient from west to east across the Rocky Mountains, due to the rain shadow effect of the Main Ranges, suggests that some areas in the Front Ranges and Foothills may have been ice-free in the Wisconsinan. Bayrock and Reimchen (1980) mapped the distribution of discontinuous, deeply weathered, diamictons interpreted as tills of probable pre-Wisconsinan age in the region. These deposits occur in upland areas between major river valleys (Fig. 5-4), above the Athabasca Valley till. They show leaching of carbonates to depths of more than 1.5 metres, occur above 1500 metres and are associated with Paleozoic erratics at elevations up to 2600 metres. Towards the southeast and at lower elevations near major valleys, the depth of leaching decreases to about 0.5 to 1.5 metres (Fig. 5-4) and is interpreted to reflect a possible, less extensive, intermediate aged (pre-Late Wisconsinan) advance (Bayrock and Reimchen, 1980). These deposits have a silty-sand matrix and a high proportion of locally derived clasts (mainly quartzites, sandstones, shales, siltstones and carbonates).

Similar deposits, showing deep weathering (leaching of carbonates and oxidation) were recognized by Boydell (1978) in the Foothills in the North Saskatchewan River area (Fig. 5-4). These diamictons are sandy, thin (less than 3 metres thick) and locally show a strong calcite cementation. Clasts are derived mainly from the Front Ranges and include very well rounded cherts and quartzites and weathered carbonates. Boydell (1978) inferred that the diamictons were deposited before the Athabasca valley till, possibly during the Early Wisconsinan or an older advance, on the basis of their deep weathering, fragmentary distribution and uniform morphologic expression. This interpretation was supported by stratigraphic and geomorphic evidence showing that the diamictons are locally overlain by younger Rocky Mountain tills and were deposited by glaciers that moved southeasterly, rising to relatively high elevations along the east flank of the Rockies.

In contrast, Kvill (1984) concluded that tills occurring in upland areas in the Foothills between major valleys were probably deposited during the Late Wisconsinan. His conclusion was based on qualitative data from the upland areas such as shallow soil profiles, thin weathering rinds on clasts, morphologic 'youthfulness' of landforms such as eskers, the distribution of lateral and hummocky moraines and the Fairfax Lake data.

Other possible evidence for a more extensive pre-Late Wisconsinan glaciation in the study area, includes the presence of glacial erratics at elevations well above the upper limit of ice in the Late Wisconsinan. In the Jasper area, quartzitic erratics from the Main Ranges occur on limestone dominated, Front Range mountains at elevations of at least 2380 metres (Plate 5-4c), much higher than the upper limit in adjacent areas of metamorphic rocks of the Athabasca Valley Erratics Train. The erratics locally exhibit pronounced weathering characteristics, such as deeply exfoliated surfaces and a high degree of rounding. In places, they occur as isolated rocks on flat surfaces and associated glacial deposits, if formerly present, have been removed, presumably by weathering processes. Weathered quartzite and conglomerate erratics resting on Paleozoic limestone were also found by Boydell (1978) on the summits of Front Range mountains in the North Saskatchewan River area. Erratics in the Brazeau River area occur up to elevations of 2075 metres (Kvill, 1984). Roed (1975) attributed quartzite erratics occurring at relatively high elevations east of the mapped limit of Athabasca valley till in the Edson region, to an early mountain glaciation.

In summary, although ice-free conditions probably existed at high elevations in parts of the Rocky Mountain Front Ranges and Foothills during the Late Wisconsinan, the extent of these refugia is not known. Similarly, although the geologic data suggest that a glacial event, more extensive than that indicated by the mapped distribution of Late Wisconsinan Athabasca Valley till (Fig. 5-4), occurred in the region, the evidence is inconclusive. The upper limit and timing of this event are also unclear but an Early Wisconsinan, or probably a much older, age is considered most likely. The possibility that deeply weathered tills and high level erratics in the region were deposited in the early part of the Late Wisconsinan seems unlikely, as geologic and geochronologic data from other parts of the Cordillera indicate that the maximum extent of the last Cordilleran ice sheet was not reached until the latter part of the Late Wisconsinan (Clague, 1981, 1989; Fulton, 1984).

SUMMARY AND CONCLUSIONS

The extent of Late Wisconsinan glaciers in the "ice-free" corridor region has been debated for many years. A major cause of the disparate views on the region was the use of basal radiocarbon dates on non-woody organics from bog and lake cores that were later found to be contaminated by radiogenically dead carbonate or reworked organics. Although these problems are well documented, Late Wisconsinan dates on organics from basal bog and lake sediments have led to several misinterpretations and subsequent re-interpretations on the timing and extent of glaciation in the region and have resulted in the perpetuation of models suggesting that large parts of the corridor were not glaciated during the Late Wisconsinan. The subtill chronologic data presented in this chapter do not support these models. It is recommended that future interpretations on the extent of Late Wisconsinan glaciers in the corridor region, that rely mainly on supratill dates of non-woody organics, be viewed with skepticism.

Lithostratigraphic data, with chronologic control provided by subtill radiocarbon dates on wood, indicate that the following sequence of events occurred in the Athabasca Valley in the Jasper region during the Late Quaternary. Non-glacial conditions existed throughout most of the Middle Wisconsinan, with alluvial fan sediments beginning to accumulate along the sides of the Athabasca River valley before 48 ka. A gradual transition from fluvial to glaciofluvial sedimentation occurred after 29 ka followed by the advance of glaciers down the Athabasca valley from source areas in the Rocky Mountains east and possibly west of the Continental Divide. Cordilleran ice from west of the Divide began accumulating in the Rocky Mountain Trench west of Jasper after 21.5 ka and subsequently spilled across the Yellowhead Pass into Alberta. The distribution of supraglacial metamorphic erratics derived from British Columbia indicates that these Cordilleran glaciers flowed through the Jasper region down the Athabasca valley, and out into the "ice-free" corridor area east of the Rocky Mountain Foothills. Cordilleran glaciers coalesced with Laurentide ice and were deflected southeasterly along the mountain front, depositing the Athabasca valley till over the entire area from the Athabasca River to south of the North Saskatchewan River valley.

Geomorphologic, lithologic, stratigraphic and chronologic data support this model of Late Wisconsinan glaciation. In particular, the distribution of surface metamorphic rocks of the Athabasca Valley Erratics Train and quartzitic erratics of the Foothills Erratics Train along a southeasterly trending belt indicates that the same Late Wisconsinan Cordilleran glaciers that flowed through the Jasper region extended down the Athabasca valley into the "ice-free" corridor area east of the Rocky Mountain Foothills where they were deflected southeasterly by Laurentide ice. The regional southeasterly trend of fluting and drumlin orientations in the Athabasca valley till also suggests ice deflection as a northeasterly trend would be expected if glaciers were allowed to flow unconfined from the mountains down the regional topographic slope to the northeast. The presence of a stratigraphically and morphologically continuous till unit in this region, showing a similar degree of landform modification by non-glacial processes, similar soil profile development and weathering depths and no end moraines or related geomorphic boundaries, suggests that the unit formed during one glacial event. Evidence that Athabasca and Laurentide tills were deposited by confluent glaciers is also indicated by a lateral contact that is lithologically and geomorphologically gradational. Radiocarbon dates from adjacent areas support this interpretation, especially those dates from Watino and Edmonton indicating that the only Laurentide glaciation to effect the corridor region in west-central Alberta occurred during the Late Wisconsinan. Sub-till wood dates from the Rocky Mountain Trench, and dates on late-glacial gastropods and bison bone, are also compatible with this interpretation.

The presence of deeply leached and oxidized tills and weathered erratics at high elevations in the Rocky Mountain Front Ranges and Foothills suggests that an even more extensive glacial event effected the region in pre-Late Wisconsinan times, although the possibility that these features can be attributed to a more extensive ice cover in the Late Wisconsinan has not been totally discounted. These data strongly suggest that glaciers in the Athabasca valley extended well beyond the Rockies to coalesce with Laurentide glaciers during the Late Wisconsinan and possibly during at least one other earlier event. It is concluded, therefore, that the 'ice-free' corridor in west central Alberta was not ice-free throughout the Late Wisconsinan period and paleoenvironmental models incompatible with ice occlusion may have to be reevaluated.

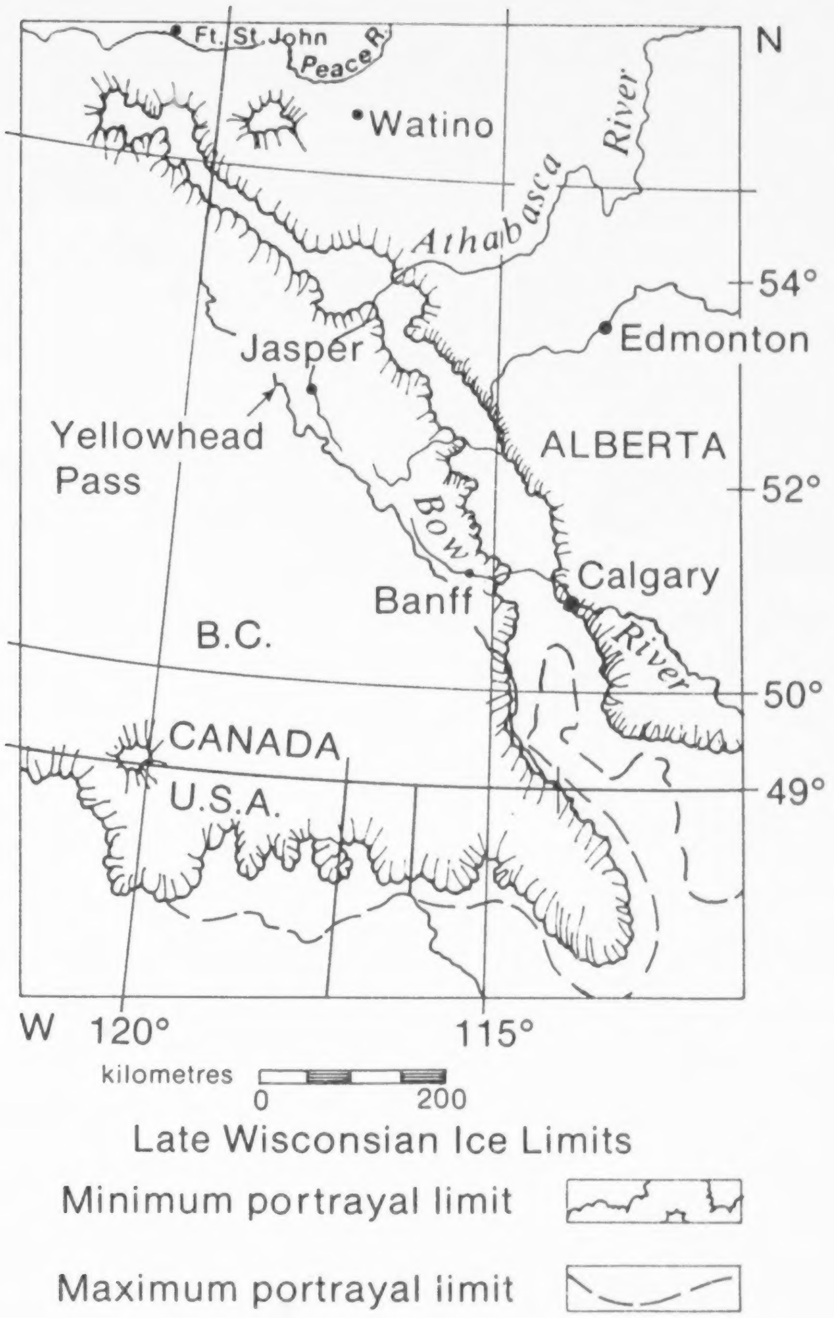
Ft. St. John
Peace R.
N
Watino
River
Athabasca
54°
Edmonton
Jasper
Yellowhead Pass
ALBERTA
52°
Bow
Calgary
Banff
B.C.
River
50°
CANADA
U.S.A.
49°
W
120°
115°
kilometres
0
200
Late Wisconsian Ice Limits
Minimum portrayal limit
Maximum portrayal limit

Fig. 5-1. Minimum and maximum portrayal limits of Late Wisconsinan glaciers in the "ice-free" corridor region in western Alberta and northeastern British Columbia (after Prest, 1984). Although ice occlusion in the Peace River area is depicted in the minimum portrayal model, other researchers propose ice-free conditions along the entire corridor region (see text).

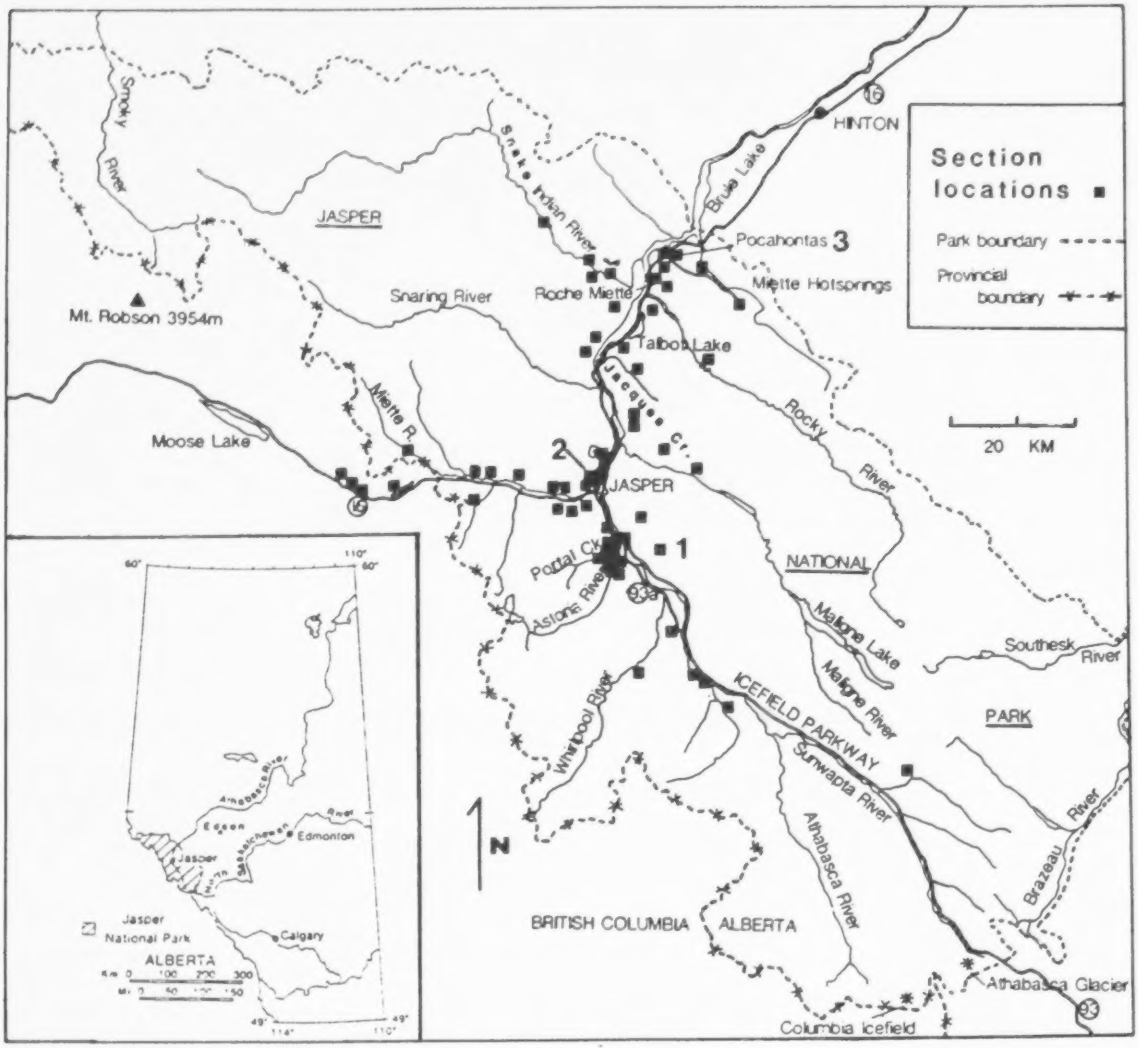
Section locations ■
Park boundary
Provincial boundary
20 KM
N
HINTON
16
Brule Lake
Pocahontas 3
Miette Hotsprings
Roche Miette
Snake Indian River
Smoky River
JASPER
Mt. Robson 3954m
Snaring River
Talbot Lake
Jacques Cr.
Rocky River
Moose Lake
Miette R.
2
JASPER
1
Portal Ck.
Astoria River
93a
NATIONAL
Maligne Lake
Maligne River
Southesk River
PARK
ICEFIELD PARKWAY
Sunwapta River
Whirlpool River
Athabasca River
Brazeau River
BRITISH COLUMBIA
ALBERTA
Athabasca Glacier
93
Columbia Icefield
60°
110°
Athabasca River
Edson
Jasper
North Saskatchewan River
Edmonton
Calgary
Jasper National Park
ALBERTA
Km 0 100 200 300
Mi 0 50 100 150
49°
114°

Fig. 5-2 Location map of place names and measured lithostratigraphic sections in the Jasper National Park region. Numbers refer to sites with chronologic control provided by radiocarbon dates discussed in the text: 1 - Amber Mountain, 2 - Jasper townsite, 3 - Pocahontas.

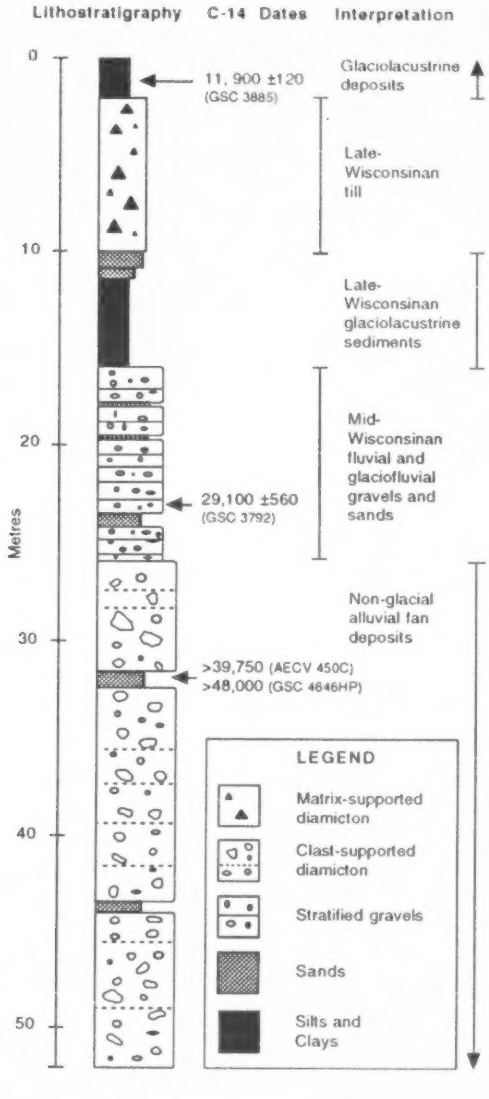
Lithostratigraphy
C-14 Dates
Interpretation
0
10
20
30
40
50
Metres
11, 900 ±120
(GSC 3885)
29,100 ±560
(GSC 3792)
>39,750 (AECV 450C)
>48,000 (GSC 4646HP)
Glaciolacustrine deposits
Late-Wisconsinan till
Late-Wisconsinan glaciolacustrine sediments
Mid-Wisconsinan fluvial and glaciofluvial gravels and sands
Non-glacial alluvial fan deposits
LEGEND
Matrix-supported diamicton
Clast-supported diamicton
Stratified gravels
Sands
Silts and Clays

Fig. 5-3 Composite stratigraphic column of Pleistocene deposits in the Athabasca Valley in Jasper National Park showing stratigraphic position of radiocarbon dates and geologic interpretation of major units.

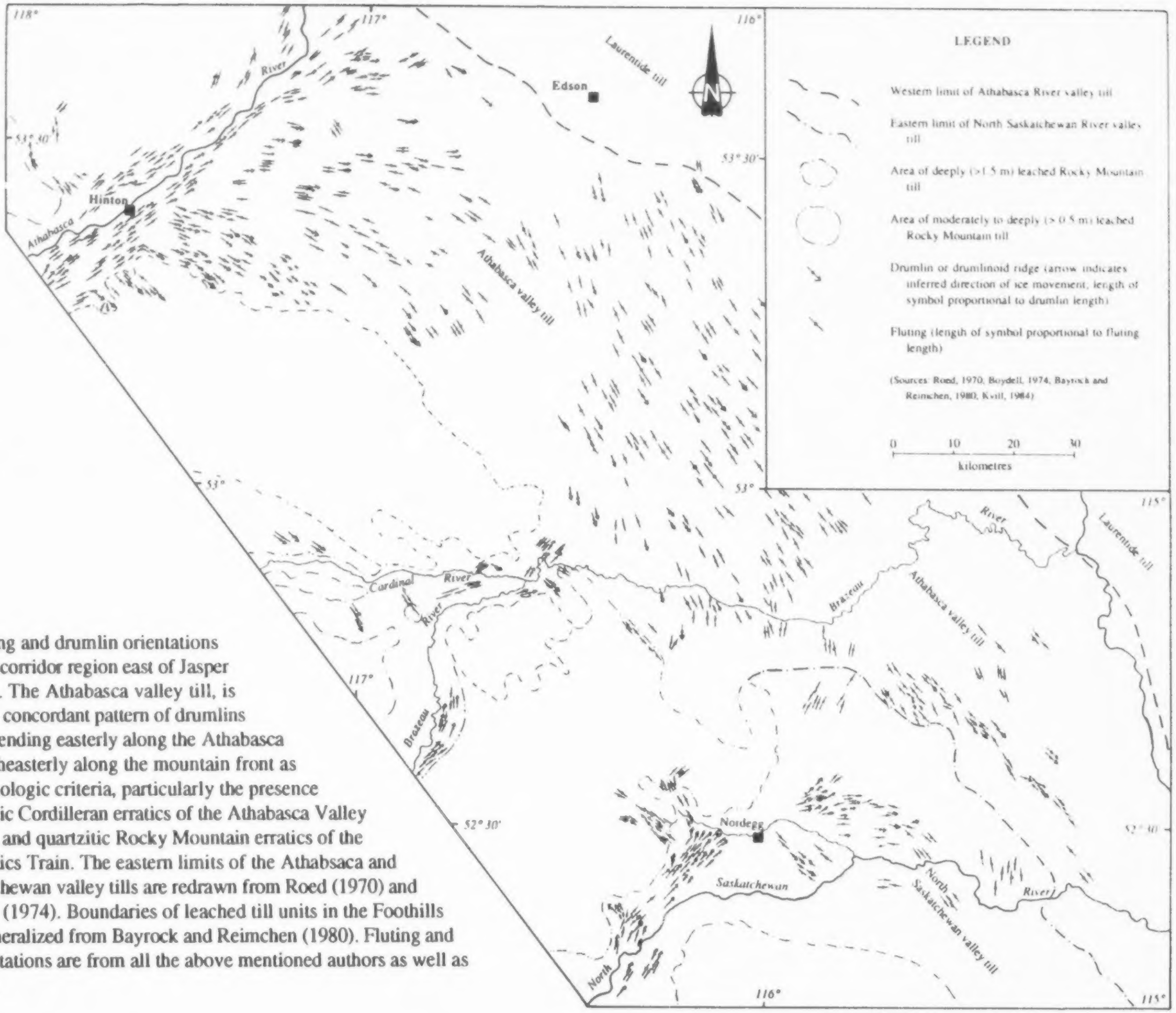
118°
117°
116°
115°
53° 30'
53°
52° 30'
River
Athabasca
Hinton
Edson
Laurentide till
N
Athabasca valley till
Cardinal
River
River
Brazeau
Brazeau
River
Athabasca valley till
Laurentide till
Nordegg
Saskatchewan
North
North
Saskatchewan valley till
River
LEGEND
Western limit of Athabasca River valley till
Eastern limit of North Saskatchewan River valley till
Area of deeply (>1.5 m) leached Rocky Mountain till
Area of moderately to deeply (> 0.5 m) leached Rocky Mountain till
Drumlin or drumlinoid ridge (arrow indicates inferred direction of ice movement, length of symbol proportional to drumlin length)
Fluting (length of symbol proportional to fluting length)
(Sources: Roed, 1970, Boydell, 1974, Bayrock and Reimchen, 1980, Kvill, 1984)
0
10
20
30
kilometres

Fig. 5-4 Fluting and drumlin orientations in the ice-free corridor region east of Jasper National Park. The Athabasca valley till, is defined by the concordant pattern of drumlins and flutings trending easterly along the Athabasca valley to southeasterly along the mountain front as well as by lithologic criteria, particularly the presence of metamorphic Cordilleran erratics of the Athabasca Valley Erratics Train and quartzitic Rocky Mountain erratics of the Foothills Erratics Train. The eastern limits of the Athabsaca and North Saskatchewan valley tills are redrawn from Roed (1970) and Boydell *et al.* (1974). Boundaries of leached till units in the Foothills region are generalized from Bayrock and Reimchen (1980). Fluting and drumlin orientations are from all the above mentioned authors as well as Kvill (1984).

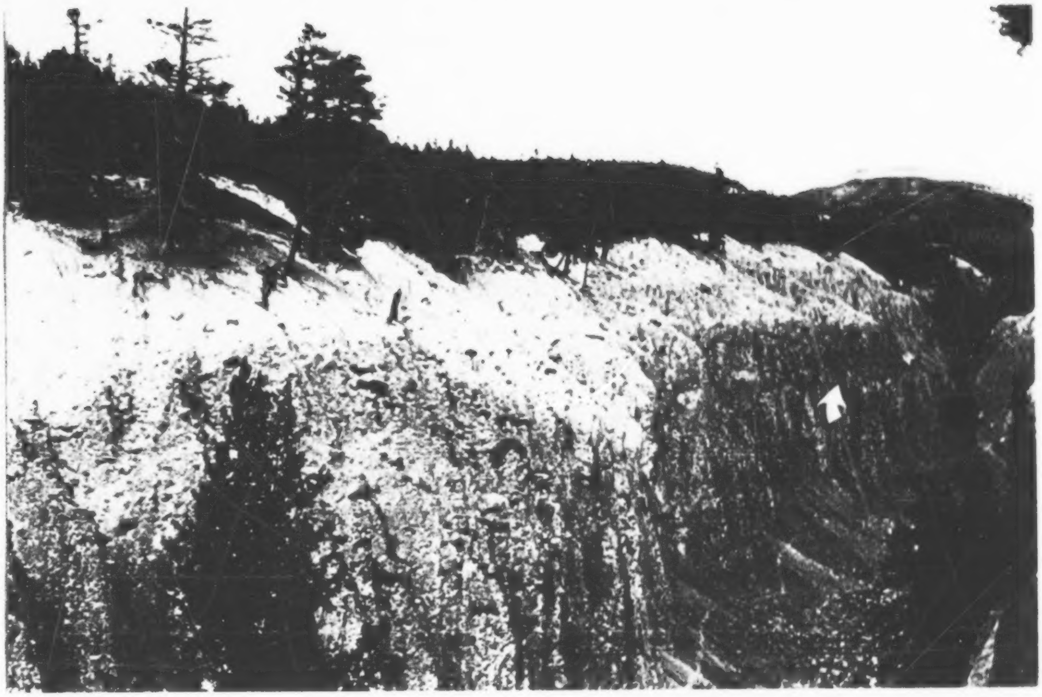

PLATE 5-1 Paleoalluvial fan, debris-flow deposits unconformably overlain by matrix-supported, glacial deposits at Amber Mountain (Location 1 on Fig. 5-2). Note the crude bedding, coarse grain-size and overall poor sorting of the alluvial fan deposits. Wood fragments yielding infinite radiocarbon dates were recovered from interbedded sand and gravel strata of fluvial origin at level marked by arrow.

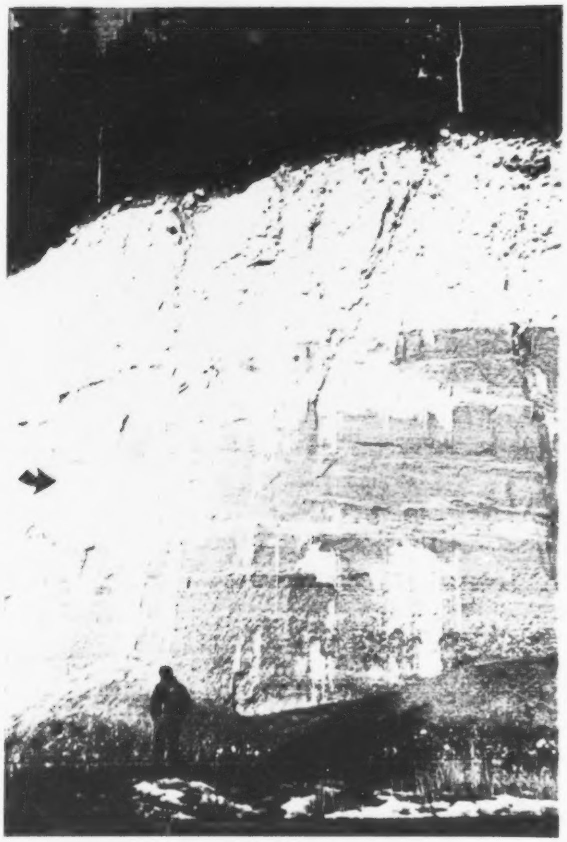

PLATE 5-2 Late Wisconsinan till erosionally overlying Middle Wisconsinan, fluvial sands and gravels near Jasper townsite (Location 2 on Fig. 5-2). Radiocarbon dating of wood (see Plate 5-3) from the fine gravel and sand unit in the centre of the plate (arrowed) indicates that deposition occurred about 29 ka .

PLATE 5-3 Sample of wood fragments deposited with sands and gravels near Jasper townsite (see Plate 5-2). The wood yielded a radiocarbon date of 29,100 ± 560 years B.P. (GSC-3792). The angular and fragile character of the fragments suggests that they are not reworked from older deposits. Scale is in centimetres.

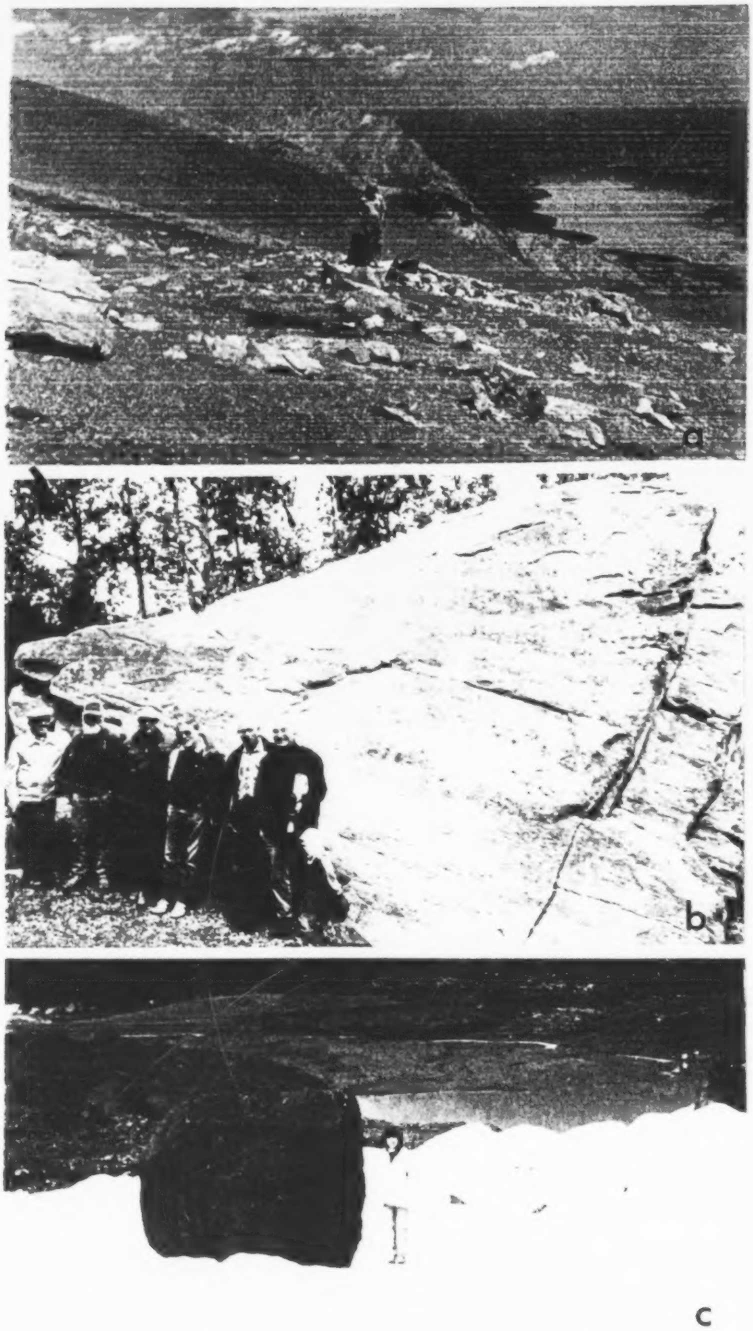

PLATE 5-4 Supraglacial erratics associated with Athabasca valley glaciers a) Angular quartzitic Gog Group erratics (arrowed) amidst locally-derived, limestone boulders near the summit of Roche Miette in the Rocky Mountain Front Ranges. The Gog Group erratics come from the Rocky Mountain Main Ranges at least 25 km distant. Geologist (A. Gambier) for scale. b) Gog Group erratic associated with Athabasca till in the Rocky Mountain House area. The erratic was investigated during the 1989 Canadian Quaternary Association Field Excursion (Levson *et al.*, 1989); participants shown from left to right are Drs. A. Dreimanis, W. Mahaney, V. Prest, N. Rutter, A. Stalker and R. Klassen. c) Modern supraglacial erratic on the surface of the Athabasca Glacier with Dr. A. Duk-Rodkin for scale.

PLATE 5-5 Ice-marginal, glaciolacustrine silts and clays overlying glaciofluvial gravels in the Pocahontas area (Site 3, Fig. 5-2). A radiocarbon date of 11,900 ± 120 years B.P. (GSC - 3885) on gastropod shells within the silt and clay unit (arrowed) provides a minimum date for retreat of Late Wisconsinan glaciers in the Athabasca River valley. D-handle shovel at section base is one metre long.

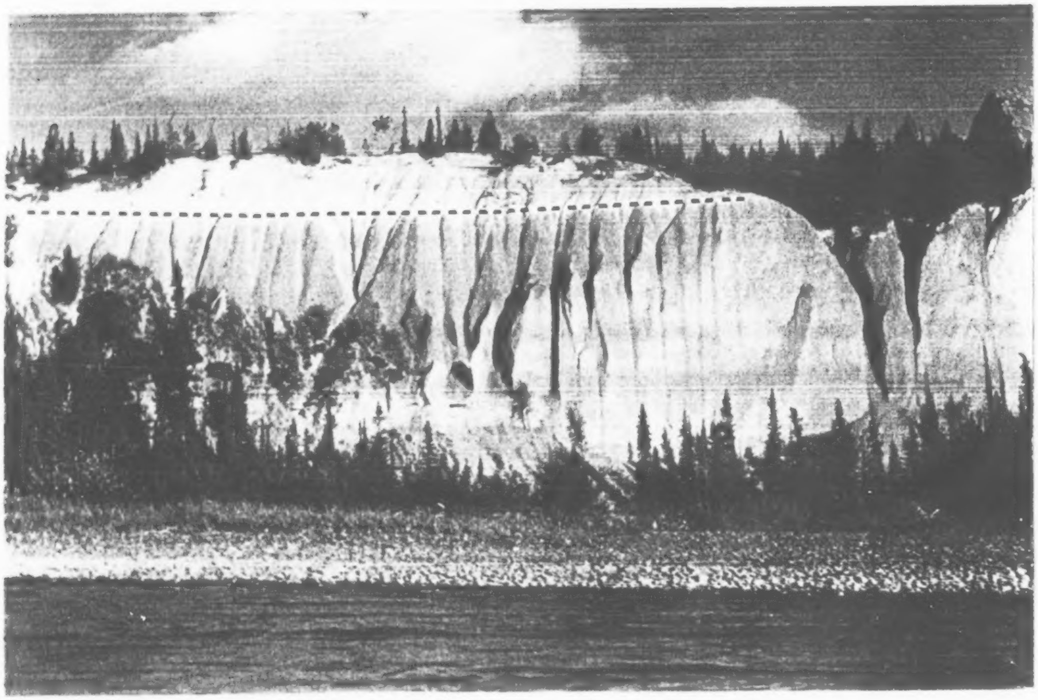

PLATE 5-6. Matrix-supported, glacial diamicton (light colored surface unit) overlying stratified sands and gravels along the North Saskatchewan River valley in the Rocky Mountain Front Ranges. The stratigraphic sequence is similar to that in the Athabasca River valley where the surface till is related to a single Late Wisconsinan glaciation.

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CHAPTER 6 - SUMMARY AND CONCLUSIONS

Few studies of Late Quaternary deposits have been conducted in the Jasper National Park region of the central Canadian Rocky Mountains. This thesis represents a compilation of four investigations on different aspects of the Pleistocene stratigraphy of the region and also has implications for understanding the late Quaternary history of surrounding areas. Six informal lithostratigraphic units are recognized from newly mapped exposures along the Athabasca River and several of its tributaries in Jasper National Park. Limited chronologic control on lithostratigraphic units in the region is provided by radiocarbon dates on wood and gastropod shells at a few sites.

The oldest unit in the study area consists mainly of locally derived, coarse-grained gravel and gravelly diamicton interpreted as Late Pleistocene alluvial fan deposits. This unit unconformably overlies Precambrian and Paleozoic clastic and carbonate rocks and stratigraphically underlies Late Wisconsinan glacigenic deposits. The paleofan successions dominantly contain clast-supported diamicton and poorly sorted gravel. Unlike arid region fans, matrix-supported diamicton is relatively uncommon. These sediments were deposited by a broad spectrum of flow types intermediate between true grain flows, cohesive flows and normal stream flows including noncohesive debris flows, hyperconcentrated flood flows and density-modified grain flows. Intermediate flow types are far more common than end members and gradational changes between them are common reflecting flow transformations due to changes in the relative concentration of the three components of the flows (water, fines and clasts) typical of alluvial fan environments. The abundance of clast-supported flows dominated by inertial forces suggests that flow velocities were relatively high in comparison with viscous debris flows that characterize many arid region fans.

Sedimentation patterns on modern alluvial fans in the area are similar to that inferred for the paleofans. Large volume, episodic, unchannelled, sediment-gravity flows are responsible for the bulk of the deposits and intermittent fluvial flows result mainly in winnowing of fines from older deposits and deposition of small sand and gravel lenses. Deposition on the paleofans occurred in a cool temperate, alpine to subalpine paleo-environment similar to that of modern fans in the region. Although debris flows are triggered by storms or sudden snowmelt events, the bedrock geology of the fan source areas exerts the strongest controls on sedimentation patterns by influencing susceptibility to debris slide initiation, drainage basin shape (and hence discharge flashiness) and availability of large clasts and matrix fines. Paleofans in the Main Ranges with mainly gently dipping, clastic source rocks are dominated by thick, bouldery debris flow sediments, deposited during high energy - low frequency events, whereas fans with steeply thrusted, interbedded carbonate and shale source rocks in the Front Ranges show thin, pebbly to cobbly debris flow beds (deposited during relatively high frequency - lower energy events) with interbedded flood flow and normal stream flow deposits.

A stratified gravel and sand sequence of inferred glaciofluvial origin gradationally overlies the alluvial fan sequences at most sites. The glaciofluvial gravels are distinguished from the fan deposits by their stratification, sorting, clast roundness and relatively high percentage of distally-derived clasts. Radiocarbon dates on rare wood fragments found in these deposits indicate that alluvial fan sedimentation in the Athabasca River valley began prior to 48 ka and continued throughout the Middle Wisconsinan. Braided stream deposits were accumulating in the valley near Jasper town site about 29 ka and, subsequently, Rocky Mountain and Cordilleran glaciers advanced through the area. In the early stages of glaciation, ice-dammed lakes formed in a number of valleys tributary to the Athabasca River valley. These ice-dammed lakes resulted in a unit of horizontally stratified glaciolacustrine sediments sharply overlying glaciofluvial gravels at a number of sites, particularly along tributary valleys in the Front Ranges. The formation of large ice-dammed lakes in these valleys indicates that glaciers occupied the Athabasca River valley at a time when the tributary valleys were still ice-free.

Late Wisconsinan glaciers deposited a complex sequence of glacigenic diamictons and associated deposits in the region that includes ice-marginal sediments, basal tills and supraglacial deposits. Erosional unconformities are common, the most pronounced occurring under basal tills. These complex deposits are well exposed in the Jasper town site area where sedimentation occurred mainly in the valley center in a pre-existing river channel. Lateral ice-marginal sedimentation along the inset valley walls was important during both the advance and retreat stages. Prior to and during glacial advance, horizontally stratified gravels, sands, and silts were deposited in a braided stream. Near the valley bottom, diamictons containing trough-shaped sand lenses were deposited by subsequent proglacial debris flows and intermittent fluvial activity. At higher stratigraphic positions, outwash deposits confined to the valley side are unconformably overlain by massive diamicton interpreted as subglacial till, deposited as ice began to over-ride the area. The diamicton grades towards the valley side into ice-contact debris deposited, at the margins of the expanding glacier, by water and sediment flows from the ice surface. Glaciers eventually overrode the entire area.

During deglaciation, sedimentation in the Athabasca valley was dominated by ice-marginal deposition of glaciofluvial sands and gravels and by paraglacial debris flows. Glacial lakes were limited in extent compared to during the advance phase. In the Pocahontas area, a radiocarbon date on shells from a small ice-dammed lake indicates that deglaciation began there prior to about 12 ka. West of Jasper town site, thick sequences of normally faulted, steeply dipping, sand, gravel, and clast rich diamicton were deposited in a kame terrace during deglaciation. The orientation of bedding and faults and the nature of the sediments indicate they were derived from, and deposited in contact with, a linear belt of stagnating, debris-covered ice that extended along the valley center east of Jasper. It is postulated that the stagnating ice originated from a well developed medial moraine and debris septum that formed at the confluence of the large Athabasca and Miette Valley glaciers. The knob and kettle topography which eventually developed in the valley center was partially obscured by glaciofluvial and fluvial activity. A linear chain of kettle lakes remain as evidence of the possible existence of the moraine. These observations indicate that kettle chains and hummocky topography, formed in the positions of both medial and lateral moraines, may be relatively common geomorphic features in glaciated mountain valleys.

Glaciers flowing down the Athabasca River valley during the last glacial period were fed both by local montane glaciers and by Cordilleran ice moving easterly across the Continental Divide from British Columbia, resulting in a major eastward flow into the purported 'ice free' corridor. The timing of this event has been disputed for many years and has important implications for the study of paleoenvironmental conditions and the north-south migration of biota in the corridor. Chronologic and lithostratigraphic data from the Jasper region suggest that this event took place in the Late Wisconsinan. Sub-till radiocarbon dates on wood in the Athabasca River valley indicate that non-glacial alluvial sediments began accumulating there sometime before 48 ka. A finite date of 29,100±560 B.P. (GSC 3792) on wood under till near Jasper provides a maximum age for the onset of the last glaciation in the region, although ice from west of the Rocky Mountain Trench probably didn't cross the Continental Divide until after 21.5 ka. The occurrence of supraglacial metamorphic erratics as well as the distribution and nature of mapped glacial deposits and the trend of subglacial paleoflow indicators in the region east of the Rockies suggests that Late Wisconsinan glaciers flowing from the mountains coalesced with Laurentide glaciers and were deflected southeasterly along the mountain front. An even more extensive glacial event may have effected the region in pre-Late Wisconsinan times. The model of Late Wisconsinan glaciation presented here is in agreement with all known radiocarbon data from the area and is especially attractive when supratill dates on non-woody lacustrine organics are viewed in light of possible contamination by old carbon.